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Field Trips Issue

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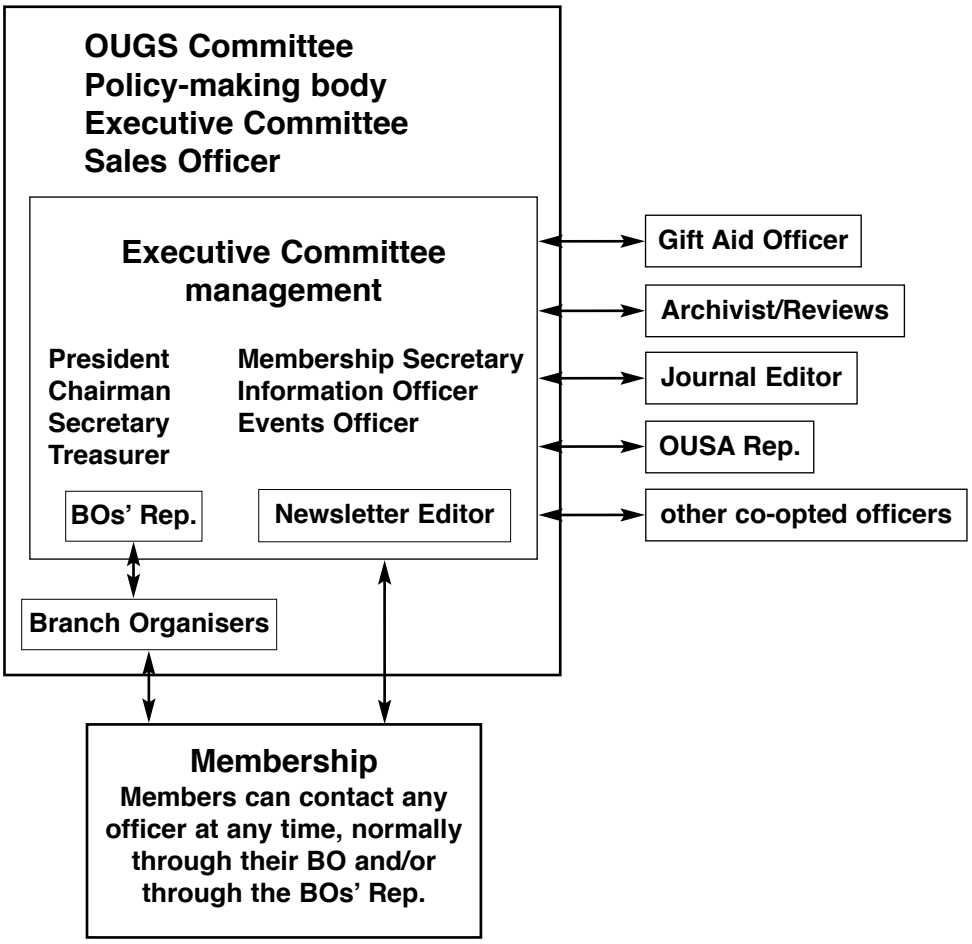

Editorial:

Dear readers,

First, an erratum:

My apology to readers. There was a mix-up in OUGSJ *33 (1) The Stirling Symposium Issue, resulting in missing information on the first page of Dave McGarvie's article 'Expedition to the Grímsvötn volcano, Iceland'.*

The article was written by Dave and based on his presentation to us at the Stirling 2011 OUGS Symposium. Essential information was omitted at the start of the article, and the separate replacement page being distributed with this issue now contains this missing information.

This issue, the second of Vol. 33 for 2012, is a 'big issue'. It comprises a splendid collection of OUGS field trip reports, starting with Dave McGarvie's OUGS Presidential trip to Iceland in 2011.

Our inveterate contributors Duncan Woodcock and John Downes have written articles on further explorations of thier favourite stomping grounds: Duncan on beachrocks in La Palma, in the Canary Islands and John on the Sawdde Gorge in the Black Mountains of Wales. Cooperative authorships have contributed articles on field trips to places around the world. There is an OUGSME report on their trip to China, an OUGS trip to the Caledonides in Norway and Sweden led by Mark Anderson of the University of Plymouth, and several field trips in Scotland — including OUGSWsx to Assynt, and two Stirling Symposium trips in the Trossachs: Tom Miller on Callender, and John Mendum on Duke's Pass and Loch Katrine).

In three further articles, by new contributors, Colin Humphrey describes a landslide in Ban Dolhelfa in Powys, Philip Hadland reports on some recently discovered potential dinosaur footprints in the Lower Greeensand near Folkestone, and Oneta Wilson and Alan Dyer contribute the first part of an analitical article on zeolites to round off the issue.

There's something for everyone here.

Our aim with the OUGSJ *is to privide members with two issues each year: one with articles from the OUGS Symposium papers (write-ups by the speakers of the lectures at the annual OUGS Symposium) and a general issue comprising OUGS field trip reports and interesting articles by members on their own geological excursions and work.*

The aim is the publish field trip articles concentrating on the geology explored by the trip, with substantial geological content, but not necessarily of academic 'peer review' calibre in terms of primary research. We would of course be more than willing to publish original research as well!

So please don't be shy. If you have gone on a great field trip and would like to share your *geological discoveries with your fellow OUGS Members, please write an article for your journal. Many articles for the* OUGSJ *are derived from brief reports in one of the branch newsletters. As Editor, I will give you as much help as you want me to.*

As always, I wish you happy reading.

— David M. Jones, OUGS Journal *Editor*

OUGS Presidential trip to Iceland — 26 August–6 September 2011

(Led by Dr Dave McGarvie, The Open Uuniversity)

Tamara Beck-Slinn, Ron Blacoe, Geoff Downer, Brian Lannigan, Caroline Lannigan, Averil Leaver, Sue Mills, Bob Morley, Cliff Probert, Lyn Relph, Bill Richardson, Jane Schollick and Averil Wootton

Introduction

The Presidential Trip to Iceland in 2011 was a special trip for all 25 participants, some of whom had never been to Iceland before (Fig. 1). Our leader, Dr Dave McGarvie (OUGS President), has done many years of research on the interactions between ice and volcanic eruptions in Iceland, so he planned the 11-day trip around the south of Iceland, where his research has been concentrated. He was able to show us many interesting unusual geological features off the beaten track; one of the highlights being a trip to see the infamous Eyjafjallajökull volcano and the effects of its 2010 glacial flood. Altogether this was a fantastic trip, with an excellent and enthusiastic leader. Many thanks to Dave and also to Chris Arkwright who organised the trip.

Figure 1 Group photograph below Eyjafjallajökull (© Dave McGarvie 2011).

Day 1 Mid Atlantic Ridge (26 August)

by Jane Schollick

We arrived at Keflavik airport on a glorious clear sunny afternoon, so Dave decided to take us on a tour of the Mid Atlantic Ridge in south-west Iceland to 'get our eye in' while the weather lasted. As Roland (our driver for the week) drove us in our 4WD bus over a vast basaltic lava field we soon began to recognise ropey pahoehoe lava. We also saw two linear, aligned ridges, looking like Toblerones, geologically known as 'tindars', which had been formed by basaltic fissure eruptions under ice. In a quarry we could see a cross-section of a tindar, showing that it was built up from a base of grey pillow lavas, over which were black and orange-brown bands of rock. Dave explained that the orange-brown rocks were palagonitised hyaloclastites, whereas the black rocks were poorly consolidated hyaloclastites.

Hyaloclastites are glassy shards of basalt formed sub-glacially (the equivalent of sub-aerial scoria or tephra). Palagonitisation is a process whereby hyaloclastite particles become welded together by chemical changes to the glass, promoted by heat. Our geological vocabulary was increasing at an alarming rate already! Our appetites whetted, we then headed for our first hotel, the rather posh Radisson Saga Hotel in Reykjavik.

Day 2 Þhingvellir, Skjaldbreiður and Hlöðufell (27 August)

by Bill Richardson

Our first stop was Þingvellir, the site of the world's first recorded parliament (AD 930) — a place I have been longing to visit since S102 introduced me to sea floor spreading. As most will be aware, Iceland is not only on the Mid Atlantic Ridge, but is also above a mantle plume. This unique combination has raised the ridge above sea level and the spreading is expressed here as an extensional normal faulted graben 8km to 9km across. Figure 2 *(overleaf)* shows how rotational movement has downthrown

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Figure 2 Western side of the Mid Atlantic Ridge at Thingvellir (© Cliff Probert 2011).

blocks that have separated from the headwall and dip towards the centre of the graben, which is filled by Lake Þingvallavatn.

The faulted rocks are basalt pahoehoe lava flows up to 2–3m thick. The headwall is in effect a cross-section through a shield volcano that erupted *c.* 9,000 years ago. Many flows show crude columnar jointing. Þingvellir is a truly spectacular location, walking in an area where the earth is splitting apart; not surprisingly, it is a popular tourist destination.

Our next locations were off the beaten track. The bus was soon bouncing alongside the shield volcano Skjaldbreiður. In Iceland the last glacial period ended *c.* 9,000 years ago. This was followed by rapid isostatic rebound, resulting in reduced pressure in the upper mantle, which in turn led to increased melting and volcanism.

In profile Skjaldbreiður has a gentle slope of *c.* 8°. On the top the outline is broken by ramparts, which are the solidified remains of the most recent lavas to be erupted. Close to the road the ropey surface of pahoehoe lava can be clearly seen. The surface is frequently broken by 'tumuli' a few metres across, where inflation of the lava flow had caused the surface crust to dome up like a blister, often followed by fracturing and collapse. Later we saw the remnants of shield volcanoes heavily eroded by ice.

Skjaldbreiður is therefore the location of a single vent that erupted low-viscosity lava less than 9,000 years ago.

We had our first view of an icecap on our way to Hlöðufell, a steep-sided mountain with a flat top known as a *tuya*, a common feature in Iceland. On the top was a thin unit made up of layers of lava. The steep slopes are a brown colour, and the lower slopes are covered in scree, moss and other debris. On approach there are many sandy brown, house-size boulders. They consist of a matrix of glassy shards, which often support angular clasts of basalt lava. On the mountain it was possible to make out foresets and other bedding features within the brown layer. A stiff walk up a stream took us to the base of Hlöðufell, where we saw that the base was made up of pillow lavas (Fig. 3).

The interpretation is that the initial eruption of Hlöðufell was under ice more than 1km thick. Basalt can melt *c.* 14 times its own mass of ice, and as the basalt entered the melt water, pillows were formed. There is a limit to how far these pillows can spread horizontally owing to the ice cover, so as the eruption continued the most efficient route for the lava to take was upward — so pillows start to build vertically. As the pillows grew and melted more water, the ice cave grew and the top of the glacier started to sag downward. Eventually the water and ice pressure dropped to a point where it was unable to contain the lava as pillows. The erupting lava started to explode violently, producing a tephra made up of tiny shards of hyaloclastite glass. This tephra slumped and flowed down the side of the mountain, producing the steep cliffs and bedding features. The tephra quickly bound together to form a solid rock by a process called palagonitisation. As the eruption continued, it melted all the way through the ice and eventually the tephra built up to the water surface. At this point the lava was erupted subaerially and started to behave as a shield volcano, building an island of pahoehoe lava surrounded by a moat of melt water. When lava reached the water it quenched and fragmented, tumbling down the sides and becoming mixed with the tephra.

Figure 3 Hlodufell basaltic tuya (© Averil Leaver 2011).

For someone whose interest in geology started by wanting to know how landscape features formed, this was a fascinating day. It showed how essentially the same earth process can form

Figure 4 Basalt columns, hotel Fluðir courtyard (© Caroline Lannigan 2011).

Figure 5 Stöng farmhouse (© Caroline Lannigan 2011).

two distinct types of mountain, depending on the environment into which the lava was erupted.

Day 3 Lava-water interactions (28 August)

by Caroline and Brian Lannigan

The main theme of the day was to study the various types of lava-water interactions, such as lava flows over riverbeds, over moist ground, and quenching by water flowing over the top. But before this, the day began in the hotel courtyard in Fluðir, just as it began to rain. The decorative basalt columns in the courtyard gave a good close-up view of their basic structure, and here Dave explained the ins and outs of column formation in cooling lava (Fig. 4).

The cooling of lava causes it to shrink by *c*. 20%, thus encouraging fractures to propagate from the exterior into the hot lava, forming typically pentagonal or hexagonal columns in basalt lavas; rhyolite lava tends to produce smaller columns of 5, 4 or 3 sides. The fracture propagation is not even, and occurs in steps. It is this repeated incremental fracturing that produces the prominent stripes across the faces of the columns, often called 'chisel marks'. We noted that the chisel marks on adjacent faces in the columns in the courtyard did not always match up. This is because the fracturing down each face occurs independently. The distance between the chisel marks is strongly related to cooling rate: the closer together they are, the faster the cooling rate. By looking at subtle textures on the surfaces of the chisel marks, referred to as 'plumose structures', it is (apparently!) possible to determine the direction in which the columns formed. When cooling is even and slow, such as at the base of lava flows, large, regular columns tend to form, and are collectively known as a 'colonnade'. At higher levels, cooling is more rapid and uneven, so the columnar jointing is much more irregular, particularly when quenching occurs by water flowing over the hot lava. This process produces a blocky or chevron entablature above the columns.

Armed with this information we set off to see *in situ* examples at Gjáin. However, our first brief stop was Stöng, the site of a Viking farmhouse excavation.

The 1104 eruption of Mount Hekla led to the devastation of about 20 farms in the Thjorsárdalur Valley, which was subsequently permanently abandoned. In 1939, Nordic archaeologists excavated the sites of some of these farms, one of which was Stöng. It was quite well preserved under thick layers of white volcanic ash, and a wooden shelter has been built to protect it from the elements (Fig. 5). From Stöng we walked a short distant to the waterfalls at Gjáin (Fig. 6) — quite a spectacular site (sight!).

Figure 6 Gjain waterfalls (© Caroline Lannigan 2011).

Here we saw exposures of a series of Holocene basaltic lava flows, which flowed into an ancient river valley, so damming the river — probably a slot canyon, Dave said (Fig. 7, *overleaf*). We saw good examples of colonnade and blocky entablature formation, and also 'peperitic' textures at the bases of some of the older lavas. Such textures are the result of lava–sediment interaction: as lava flows over wet sediment in a riverbed, it picks up some of the sediment, forming peperite. This is clearly seen as pale coloured sediment intermingled with the dark lava at the base of the lava flow. Some of the group climbed to the top of the exposure and here we saw an overall picture of the river flow down the canyon cut into the lava fields above the falls.

From Gjáin we travelled a short distance to one of Iceland's many rootless cone fields at Thjorsárdalur. They used to be known as 'pseudo craters'. They are formed by the interaction of lava flow with surface or near-surface water. This interaction can

Figure 7 (left) Holocene basalt lava flows in an ancient river valley (© Caroline Lannigan 2011).

Figure 8 (below) Rootless cones at Thjorsardalur (© Allan Metcalfe 2011).

produce explosions, rupturing the lava's crust, ejecting scoria first, building up a cone or crater and topped by a spatter cap. A cone's formation from scoria through to spatter reflects decreasing amounts of water as the eruption progresses, suggesting a limited water supply. Simple calculations have shown that the amount of material in a rootless cone is more than the underlying vent mass, which leads to the interpretation that extra material must have come from lava tubes (Fig. 8).

From here we travelled to a double waterfall called Hjálparfoss, where there are more exposures of lava–water interaction. This waterfall flows over and into a *c*. 3,500-year-old lava flow. As with Gjáin, the lava flow had filled an ancient river canyon and dammed the river, such that when the water level rose high enough to flow over the lava, it penetrated the lava from above, producing a series of fractures. The uppermost lava

Figure 9 Chevron entablature in basalt at Hjálparfoss (© Caroline Lannigan 2011).

Figure 10 Pseudopillow fracture at Hjálparfoss (© Caroline Lannigan 2011).

here is the same as the lava at Gjáin. We saw excellent examples of colonnade and chevron entablature formation and also a pseudopillow fracture (Figs 9 and 10). These features are caused by water penetrating down through a fracture in still ductile lava, producing a temperature gradient out from the fracture, which in turn causes smaller arcuate fractures to form perpendicular to the main fracture.

The final visit of the day was to a modern reconstruction of a Viking farmhouse, based primarily on the excavated ruins of the Stöng farm. The group's reaction was mixed, the highlight for some being an apparently typical 'Viking tractor / lawn mower' (wood burning?) close by!

Day 4 Dómadalshraun, Ljótipollur and Laugahraun (29 August)

by Averil Leaver

As we drove north-east around a cloud-shrouded Hekla, Dave gave a brief description of its history and significance. Strombolian and Plinian eruptions along a fissure have created a ridge-shaped stratovolcano associated with extension in the transform zone between east and west rift zones. We crossed the head of the a'a' lava flow from the 1847 eruption at Lambarfitjarhraun.

Within view of the Torfajökull volcanic system we made a brief stop at Dómadalshraun to examine evidence of the interaction and mixing of basalt and rhyolite magmas, a result of basalt movement laterally in dykes from the rift zone fissures to the south-west toward Torfajökull. This eruption, dated AD 150 by tephrochronology, had produced pumice from an initial Plinian eruption and hybrid magma, and we examined blocks containing wispy contacts between differently coloured areas, flow-banding and phenocrysts that had grown in the magma chamber before the eruption.

At Ljótipollur we observed the maar (crater) produced when a basalt dyke transporting magma from Bárðarbunga toward Torfajökull encountered an aquifer in the bedrock. The explosion produced by hot magma in contact with water in a confined zone ripped out a 25m deep pool and draped the surrounding slopes with spatter (Fig. 11, *overleaf*).

Walking from Landmannalaugar westward, we skirted Bláknúkur, a mountain probably formed by a series of rhyolite eruptions under ice, the youngest of which dated to 125ka. Columnar and well-jointed lava bodies represent rhyolite lava flowing into the surrounding ice. In contrast to the palagonite formed from basalt lava–water interactions seen earlier in the trip, rhyolite lava–water interaction produces less-cohesive 'piles of crud', which drape the slopes (Fig. 12, *overleaf*).

As the day brightened we turned onto the Laugahraun rhyolites lava field, erupted subaerially in a warmer period, in fact just a few hundred years ago, *c*. 1477. We picked our way across to the line of fumaroles that have exploited faults (Fig. 13, *page 7*), stopping briefly to examine a ductile fracture feature on blocks on either side of the path (Fig. 14, *page 7*). The dome on the high point of the lava field had collapsed outward during the final stage of the eruption, to leave a petal-like arrangement of blocks marked in places with a Raku-style pattern that intrigued Dave — yet another question to be answered.

Figure 11 Ljotipollur maar at Ljótipollur (© Allan Metcalfe 2011).

Day 5 Hnausapollur, Námshraun and Stútshraun (30 August)

by Ron Blacoe

On a dry but overcast day we left our hotel at Hrauneyjar, which is one of Iceland's major centres for production of hydroelectric power. It is also a seismically active area, a cause of concern. The majority of rivers we crossed were heavy with 'glacial flour', which can cause extensive corrosion in the hydroelectric turbines. Therefore there are ponds or lakes in which the larger particles can settle out.

We visited three locations: Hnausapollur (a basaltic maar), Námshraun (a rhyolite lava flow and dome) and Stúshraun (a complex basaltic lava flow with scoria cone).

On the way to the first location we had a brief stop to view the Eastern Rift Zone from a north-westerly direction. From north to south we could see a range of mountains that included Thoristindur, Loðmundur and Hekla. We discussed their composition and Dave pointed out that the most north-easterly ones were basaltic, while to the south-west they tended to be rhyolite. These basaltic hills, and beyond to the north-east, are the source of the basalts that intrude the rhyolite, causing rhyolitic eruptions contaminated to varying degrees with basalt.

Like other maars, Hnausapollur basaltic maar formed by the deep interaction of basaltic magmas associated with fissure eruptions and aquifers. This maar, *c.* 100m deep, was formed in AD 870. This is much deeper than Ljótipollur (*see* Day 4). From the maar rim we were able to view 'bulls eye rings' on the nearby hill of Grákolla, which are the exposure of different eruptions of tuff, each of which can be dated and are therefore useful for working out the chronology of the volcanic activity.

We then visited Námshraun, a subaerial blocky rhyolite lava flow, which erupted *c*. 1477. The eruption is estimated to have lasted approximately one month, and the main vent complex produced rhyolite lavas that flowed to the south-east and into lake Frostastaðavatn and north-east into the braided river plain of the Jökulgilskvísl River. We climbed this river plain flow to view the main rhyolite dome and enter it (Fig. 15, *opposite*). There is evidence of a brecciated carapace and the mechanism involving ductile/brittle interaction whereby the dome releases pulses of lava.

Figure 12 Bláknúkur rhyolite hills near Landmannalaugar (© Dave McGarvie 2011).

Figure 13 Fumaroles at Laugahraun (© Allan Metcalfe 2011).

Figure 14 Rhyolite fractures at Landmannalaugar (© Bob Morley 2011).

We retired to Landmannalaugar for a well-earned lunch break and some of us had a soak in the hot springs running from beneath the main lava flow (Fig. 16).

A short drive took us to Stútshraun a subaerial basaltic pahoehoe lava flow. There were at least two active vents, one with a well-preserved scoria cone, which is unusual, as scoria cones are inherently unstable and erode rapidly. There are also 'hornitos' (chimneys, from the Spanish for 'little horns'), as well as circular pit-like collapses in the lava surface (Fig. 17, *overleaf*). These pits caused a lot of discussion, but no conclusion was reached. The lava had also experienced extension at 90° to its direction of flow, exhibiting both ductile and brittle fracturing. There is also evidence of lava levees, and Dave told us that the lava is contaminated with silicic material, presumably from an older and buried rhyolite unit or units. Once we 'got our eye in' we could see numerous lava tubes within the flow, which exhibited stalactite features.

All in all a very unusual and complex basaltic lava flow.

Day 6 Eldgjá gorge and Skaftá River (31 August)

by Jane Schollick

We travelled south-east from Hrauneyjar to Kirkjubaejarklaustur, during deteriorating weather. On the way we stopped to see a dramatic multiple waterfall from a leaky hydroelectric power station dam. Water leaked along the contact between the impermeable pillow lavas and the porous Holocene basalt above them (Fig. 18, *overleaf*).

After passing south through the rhyolite hills around Landmannalaugar, we stopped at the misty Eldgjá gorge to have lunch on the bus, as it was raining hard by now (Fig. 19, *page 9*).

Eldgjá is the site of the source of Iceland's largest known historic eruption in AD 934–940, a huge flood basalt eruption that produced *c*. 20km^3 of basalt (even more than the Laki flood basalt of 1783). While walking along the river gorge for *c*. 1.5km we examined two huge lumps of basalt labelled as recent rockfalls, a 3m high mound of scoria and spatter welded together and a large hornito. Dave told us that this u-shaped gorge looks glacial but is really an old graben, utilised by the basalt as an easy route to the surface.

There are two theories about the origin of this flood basalt. First, it could be part of a fissure swarm linked to Katla, as the Eldgjá basalt is linked chemically to Katla's basalts. Some consider this explanation unlikely, as it is impossible to get this huge volume of lava from one volcano. Second, it could be linked to the mantle plume, which is not near Katla. This explanation would produce enough magma, but doesn't fit chemically.

Clearly, more research is needed.

Figure 15 Námshraun rhyolite dome (© Shirley Blacoe 2011).

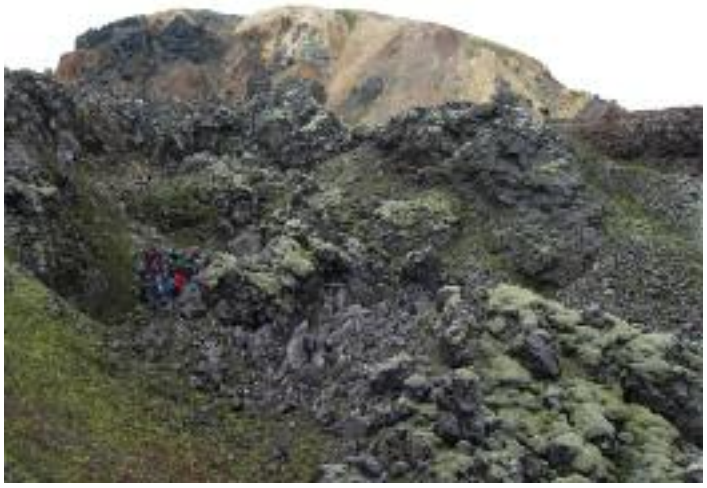

Figure 16 Group in hot pool at Landmannalaugar (© Sue Edwards 2011).

Figure 17 Inspecting a 'hornito' at Stútshraun, scoria cone in background (© Averil Leaver 2011).

Some miles farther on, we stopped at the Skaftá River, which is subject to frequent glacial outburst floods, as it drains an active geothermal area under the Vatnajökull ice cap (Fig. 20, *opposite, bottom left*). Fortunately the river was not in flood, so we were able to examine the riverbed safely. Here there were exposures of the Eldgjá lava, in the form of small irregular columns, in all orientations. This indicates that there must have been an additional coolant, probably river water, as the Eldgjá lava would certainly have exploited river channels in this area. We also observed pseudo-pillow fractures, long curving fractures with tiny straight fractures coming off at right angles — giving an appearance like millipedes! The curving fracture was propagated through ductile lava before it had solidified, and the straight brittle fractures appeared slightly later.

Day 7 Lómagnúpur, Skaftafell and Hvalnesfjall–Krossanesfjall (1 September)

by Cliff Probert and Bob Morley

The rain falls mainly on the net-veined complex: greeted by another wet and windy day we changed our itinerary to take in sites farthest east, to save driving for a later, hopefully better, day. We headed east toward Skaftafell.

We stopped briefly to observe the old sea cliffs at Lómagnúpur, which is thought to be the highest (*c*. 600m) inland cliff in Iceland. It is composed of layers of palagonitised tephra and basalt sheets from the Siða Group.

Our route crossed several bridges over glacial rivers and outwashes including the Skeiðará River, previously Iceland's largest glacial river, but now only a shadow of its former size because its headwaters have been captured through flowing along the glacier margin into another river. We stopped in a lay-by to see the remnants of a bridge at Skeioararsanður destroyed by the 1996 flood from the Gjálp eruption. From this point we were also able to see three of the valley glaciers that emanate from the vast Vatnajökull icecap to the north (Fig. 21, *opposite, bottom right*).

From here we moved on to Svínafellsjökull and for a short while we had some respite from the rain. At the end of a rough track were able to overlook the end of a valley glacier with its massive landscape of crevasses and dirty ice owing to the brownish ash fall from the recent Grímsvötn eruption. The maximum extent of the glacier is thought to have occurred *c*. 1875–80, and since then the glacier has been in retreat and is now thought to be retreating at *c*. 120m per year.

The path we walked on lay well above the ice, but the presence of striations on many surfaces and rounded boulders indicated that ice

Figure 18 Waterfalls from leaky HEP dam (© Dave McGarvie 2011).

once reached much higher up the valley walls. The rocks in this area are dated 4–2Ma and pre-date the formation of the younger Öraefajökull stratovolcano (*c*. 700ka old). In the valley walls above the glacier we observed several sills and dykes, which must have come from a central volcano.

The British geologist George Walker, one of the most influential volcanologists in the world, gained a reputation for his work on zeolite minerals that grow in buried basalt lavas and used zeolites to decipher the growth of the lava pile in eastern Iceland. He recognised that certain zeolites form within specific pressure ranges, and that by careful analysis of zeolites in eastern Iceland the depth of burial of the lava pile could be estimated. This was groundbreaking work.*

Figure 19 Eldgja river and gorge (© Sue Edwards 2011).

We lunched at Skaftafell visitor centre, where many of us took the opportunity to watch a short film about the Skeiðará glacial outburst floods (jökulhlaup) from the 1996 Gjálp eruption, and to study the exhibition summarising the 1362 (rhyolite) and 1727 (basalt) eruptions from Öraefajökull.

After lunch we continued east to the glacial lagoon of Jokulsárlón, but with deteriorating weather it was decided to postpone the boat trip on the lagoon to another day and limit our stay to photographing some of the icebergs and seals in the lagoon (Fig. 22, *overleaf*).

There then followed a long drive to a net-vein complex in the Hvalnesfjall-Krossanesfjall area, east of Hofn. This is a *c*. 8Ma fossil magma chamber (*c*. 2km^2) that has become exposed due to erosion of the overlying extrusive rocks, and is a mixture of basalt and rhyolite injected 4–5km below the surface. This is considered to be a granophyre formed from the partial mixing of hot mafic magma as it intruded into colder acidic magma (rhyolite).

Figure 20 The Skaftá River lava flow (© Bob Morley 2011).

Elsewhere there were examples of dark grey gabbro-dolerite. The complex exhibited both micro and macro partial mixing with some boundaries being 'freckled', while others were straight and jagged. The partial mixing at the edges appeared to be mostly liquid but there is also evidence of brittle fracturing, possibly because the basalt was injected into relatively cold rhyolite. With a hand lens, pale cream-pink feldspars, black pyroxenes up to 3mm and quartz crystals can be seen here (Fig. 23, *overleaf*).

Day 8 Icelandic rhyolite, Jokulsárlón and Hvalvorðurgil (2 September)

by Averil Wootton

Foul weather again hampered our plans. We'd hope to be able to view Kálfafellsdalur, the valley where we had surveyed the moraines in 1971, a 'spectacular example' of classic alpine landscape, but all four times we passed its entrance the view was obscured.

So we started in the comfort of the hotel, where Dave held an informal tutorial on the mystery of Icelandic rhyolite — in short, why is there so much of it? Iceland is 11% rhyolite. Fractional crystallisation can only be part of the answer, as melted basalt, under laboratory conditions, yields only 3–5% of its original mass as rhyolite.

To summarise a detailed account: the answer may be a case of unique geological circumstances.

Figure 21 Twisted bridge remains, Skeioararsanður (© Sue Edwards 2011).

* *See page 95, this volume — Ed.*

Figure 22 Jokulsarlon Glacial Lagoon (© Sue Edwards 2011).

Figure 23 Hvalsnes net-veined complex (© Dave McGarvie 2011).

Iceland lies on a mid-ocean ridge but in close association with a mantle plume. There are two methods of producing rhyolite in these circumstances:

1. Slow fractional crystallisation of basaltic magma provided by the spreading ridge in the thickened crust (30km thick). This will give up to a maximum of *c.* 5% rhyolite.
2. Melting of older (>3Ma) crust. As fresh hot basaltic magma travels through the crust (at 1200°C) the lowest M.Pt minerals melt, e.g. those that form rhyolite. This produces the extra rhyolite.

This situation has existed for some 70 million years. As the North Atlantic opened, an extensive volcanic province formed, of which the oldest rocks are now in Greenland and Scotland, and the youngest in Iceland. In Iceland itself, the older rocks are to the west and to the east. It is possible, however, that the plate boundary and the hot spot could separate in the future, and even now the boundary 'jumps'. The plate boundary currently lies west of the plume, which rises under Vatnajökull. Also, the plate boundary (essentially the axis of the Mid Atlantic Ridge) is moving slowly north-west, away from the plume.

Next we set off for a second look at Jokulsárlón and Dave outlined the Icelandic Ice Age as we went. Trees such as magnolia and redwood had flourished in the warm Tertiary climate, a contrast with the 25% birch cover at the time of Settlement in the late 9th century AD. Global cooling began *c*. 7.5Ma; subglacial eruptions are absent from Iceland before *c*. 4Ma, but are abundant after 2.5Ma. There have been some 10–15 glacial cycles, each between one- and two-hundred thousand years long, separated by shorter, milder interglacial periods (i.e. based on Milankovich cycles).

The cause of ice ages is still imperfectly understood, despite the claims of earlier workers, but an astronomical cause remains the most likely mechanism. The ice not only affected erupting lava and eroded the landscape, but also led to vertical movement of the land, as a result of loading and unloading as the ice formed and then melted. These isostatic adjustments influenced changes in both relative sea level and the tilt of the rocks.

At Jokulsárlón, the proglacial lake of Breiðamerkurjökull, we had the full tourist experience. We started with the stranded 'bergy bits' on the seashore below the outfall (Fig. 24). Cloud, mist, grey thundering sea and black sand created a mysterious backdrop for the gleaming lumps of ice, which appeared to be illuminated from within. Many photos later, we boarded an amphibious vehicle for a trip around the floating bergs on the lake itself. The lack of sun meant ideal conditions to see the ice at its bluest, an effect enhanced by the contrast with our orange life jackets (Fig. 25).

Our Icelandic guide explained that the ice front has retreated *c*. 7km over the last 80 years, and that the lake now has an area of 23km^2 and fills a fjord 250m deep. To me, these figures were deeply shocking, but explained why my memory had seemed faulty when we'd seen Breiðamerkurjökull the day before. I recalled it as a modest little lake in 1971, with an outfall easily spanned by a fairly short Bailey-type bridge. The ice surrounded a classic nunatak in the far distance, but that is now a promontory that divides the ice front into two parts. Judging from the information board, the lake area has tripled since I first saw it. My memory is, after all, in order, but this is poor recompense for the reality of global warming.

Finally, on to Hvalvorðurgil on Öræfajökull for 'rhyolites in the rain — close-up and personal' as Dave put it. Here, we had the authentic geological field-trip experience for those who felt up to it — a rough clamber up a sodden slope of grass and scree in the pouring rain, very intimidating to the uninitiated. It was worth it, though, to see rhyolite that must have flowed through a subglacial tunnel. This rock had a chilled margin of glassy, flow-banded rhyolite in which exquisite hand-specimen sized columns, complete with miniature chisel marks, had formed in a radiating pattern. We also glimpsed a rhyolite flow perched above us, formed where the lava ran along the edge of the ice but lacked the heat to melt its way through. Once the ice melted, the lava was left high up in an improbable position on the mountainside (Fig. 26, *overleaf*).

Figure 24 Icebergs on the beach at Jokulsárlón (© Sue Edwards 2011).

Figure 25 Ice at Jokuslárlón (© Allan Metcalfe 2011).

Day 9 Svínafell, Haálda, Thvera, Mulakvísl bridge and Reynisfjara cliffs (3 September)

by Tamara Beck-Slinn

We travelled west from near Höfn to Vík, going into the Lower Pleistocene — when small outcrops would have been big islands and mountains; and when redwoods and magnolias would have flourished here. One km in height has been eroded away, mostly in last 2.5Ma. And because of this erosion 10Ma of rocks are missing, as the oldest rocks seen in Iceland are *c.* 16Ma old. The mantle plume and mid-ocean ridge met 25Ma ago. Big, isolated lava heaps in Iceland indicate that it could have been bigger than it is now, showing that glaciers are far more erosive than weather.

We had two view stops. The first was a scenic panorama just as the mist lifted from Svínafell (Fig. 27). Looking east we could see a perched rhyolite flow, old brownish lavas, river terraces and, at 800m, a 'scab', possibly the remains of an old pyroclastic flow. Below these are subglacial basalt and a lateral moraine slanting up towards the glacier. At the foot of Kviárjökull there is a *c.* 100m high terminal moraine — the tallest in Iceland.

The second view stop was to see three kettle holes at Haálda, left behind by a jökulhaup during the 1727 eruption of Öraefajökull. 'Warm' water carried meltwater and large lumps of ice onto the lowlands, down a narrow rock channel. As we drove around Öraefajökull, we could see its steep basalt cliffs with 16 glaciers. Just before Skaftafell, a pale area on the cliff shows evidence of 3Ma lake sediments containing fossil trees and leaves.

Figure 26 Rhyolites in the rain at Hvalvorðurgil (© Averil Leaver 2011).

Figure 27 Svínafell glacier (© Sue Edwards 2011).

Figure 28 Basalt cliff at Thver (© Bob Morley 2011).

After Skaftafell, we drove through an area of younger Ice Age rocks (<2Ma old).

Once again we traversed the extensive Skeiðarársandur, passing at the western end the 600m high Lómagnúpur cliff (sheet basalts dated 0.8–2.0Ma), comprising seven mapped units. We stopped for a close look at Thvera locality, where we were slightly delayed by the traditional first day of sheep gathering. It was poignant watching the skilled horse and dogs in the drizzle. We took a steep walk up to a cliff, where there is a basalt horizon with tephra above and tillite below. It has chilled margins on both sides. Molten stringers of basalt have been injected 10m up into the *c.* 50m high tephra layers. This process needs more research to explain how and why. Brown Grímsvötn ash from the May 2011 eruption had collected in the hollows at our feet (Fig. 28).

The next stop was to see the new Mulakvísl bridge, built in a week after a 5m high wall of water and debris thundered down from Mýrdalsjökull at 3am in July 2011 and destroyed the old bridge (Fig. 29, *overleaf*)! Icelanders have an effective way of building bridges: they divert the river to one side, build the bridge on dry land and then reinstate the river. There is a lingering smell of sulphur from the river, reminding us that it drains active geothermal areas beneath the Mýrdalsjökull glacier!

The highlight of the day was the 100m high sea cliffs at Reynisfjara, a stop carefully timed for low tide because of freak waves. This is a beautiful beach with black pebbles and white breakers. Hard hats were worn, as the cliffs are continually eroding. A weird platy formation goes right in at the base of the cliff —fracturing? It forms a great contrast to the basalt columns with pipe vesicles above. Bland, very unconsolidated, tephra lies with stringers above the basalt columns, with pahoehoe lava as the topmost layer (Fig. 30, *overleaf*). Farther east (nearer the stacks) the columns are more glassy. Dave's useful tip: If there's a rumble from above and the rocks are small, run *towards* the cliff; if the rocks are large, run *away* from the cliff!

Day 10 Solheimajökull, Skógafoss and Gigjökull (4 September)

by Lyn Relph

We started the day in Dyrhólaey, where we walked up to the lighthouse so that we could look back to the Reynisfjara basalt sheets that we had visited the previous day in the pouring rain. In the bright sunshine they looked stunning with black sands and sparkling sea. Those with a head for heights walked over the natural arch to look at subaqueous basalt deposits. Sadly I was not one of them (Fig. 31, *overleaf*).

The next stop was Solheimajökull, where there is an impressive valley glacier. We approached by a track along the side of a gravelly river. This gravel is the larger particles of the glacial deposits;

Figure 29 Mulakvísl July 2011 flood damage (© Shirley Blacoe 2011).

the finer material is carried away by the river. Walking up to the glacier we passed through an area of both lateral and terminal moraines, testament to the shrinking of the glacier. Meltwater has washed gaps in the terminal moraine. Closer to the glacier are mounds of black debris, some of which still have an ice core. The glacier was not pristine white and blue, but very dirty with black streaks, due to ashfall from the Eyjafjallajökull eruption of April 2010 (Fig. 32, *opposite*). Water running along the top of the glacier washes surface debris into the hollows, forming large accumulations. When the glacier melts these accumulations are left behind as large mounds of material. There are many interesting rocks washing out of the ice, some heavily striated and some with striations in more than one direction. The differential striations are because the rocks change their orientation within the glacier, and again when washed out of it. Therefore, striations on boulders cannot be used as ice flow direction indicators.

Skógafoss waterfall was quite spectacular at the time of our visit, but during the Eyjafjallajökull 2010 eruption it ran black with ash (Fig. 33, *opposite*). Some of us had taken a liking for the sparkly bits in the rocks we found in the glacial moraines, so Dave treated us to a quick stop at a cliff to find some Ankaramite. This is a stunning rock: it has large crystals of olivine and pyroxene set in a fine dark groundmass — basically, a primitive basalt.

Eyjafjallajökull is a stratovolcano formed mostly of basalt sheet-like units that are typical of eruptions at ice-covered stratovolcanoes. During an eruption the magma melts the ice (basalt lava can melt up to 14 times its own volume of ice), and this meltwater forms channels under the ice as it flows through it. These meltwater channels are then exploited by the lava.

To end the day we visited the Gigjökull glacier, which drains Eyjafjallajökull's summit crater — from our perspective to the right of the visible peak (Fig. 34, *page 16*). At the base of the glacier, before April 2010, there had been a glacial lagoon, renowned for its beauty. This eruption, however, caused a flood over the top of the glacier and eroded huge channels in the side walls of the glacial valley, depositing a 50m depth of debris that filled the lagoon (Fig. 35, *page 16*). Blocks of ice removed from the glacier get buried in the sediment, then melt to leave

Figure 30 Basalt columns at Reynisfjara (© Cliff Probert 2011).

quicksand and large kettle holes (Fig. 36, *page 17*). The eruption removed 40% to 50% of the ice off the front of the glacier.

The devastated land is already being colonised by new plant life, quite remarkable as there is unlikely to have been a seed bank in these sediments. There is a flood warning system that gives one and a half hours warning for the authorities to close the road and for farmers to move their livestock, mainly sheep.

Figure 31 Dyrholaey (© Allan Metcalfe 2011).

Figure 32 Solheimajökull glacier (© Geoff Downer 2011).

Day 11 Hellisheiði Geothermal Plant and Seltangi (5 September)

by Geoff Downer and Sue Mills

The morning of 4 September boded variable weather, but this did not deter several of us from donning our bathing costumes and taking an early morning plunge in the hydrothermally heated swimming pool and hot tubs. There are two hot tubs, their temperatures regularly monitored for constancy by hotel staff — one between 38° and 40°C and the other between 40° and 42°C.

We planned to spend the day on the Reykjanes Peninsula, part of the Western Volcanic Zone where there are numerous hot springs and fumaroles, all signs of an active volcanic system underfoot. The peninsula is home to a north-east to south-west trending volcanic pro-vince comprising four fissure systems that have erupted subaerially in Holocene times, most recently in the 13th century. Our bus passed many lava flows and tephra deposits on the way to our first stop at the Hellisheiði Geothermal Plant (Fig. 37, *page 17*).

The geothermal power plant (one of five on the island) is at Hengill, part of an active volcanic ridge. Geothermal power provides over 25% of Iceland's energy requirements and there are plans to increase output over the coming years. At present geothermal energy provides heating and hot water to 87% of the island's buildings. What struck us all was the cleanliness of the plant, a marked contrast to a coal-fired power station. The architects and staff working there were proud of their awareness of the environment and hundreds of thousands of trees are being planted in the vicinity. The plant has also increased the flow of water into Lake Þingvallavatn and in consequence staff monitor the depth and mineral composition of the water on a regular basis.

Hellisheiði is the largest geothermal power station in Iceland. It uses steam to generate turbines to heat fresh water and has a network of 57 bore holes delving into the superhot rocks below ground. These bore holes, the deepest in Iceland, reach to a depth of 4,000m, and they can extend 1,000m laterally to utilise the 300°C geothermal heat available. The power station creates 180MW of electricity, which is *c*. 30% of the geothermal electricity produced in Iceland. A shiny turbine rotor blade on display is identical to those used in the plant and at 31.2in (0.79m) is the largest in the world.

The Icelandic government hopes to be able to provide sufficient hydro- and geothermal stations to make the nation 100% fossil-fuel free in the near future.

Figure 33 Skogafoss waterfall (© Geoff Downer).

Figure 34 Eyjafjallajökull and Gigjökull glacier (© Sue Edwards 2011).

During a break in the weather, with good visibility, we stopped for lunch amid boulders of basaltic rock along the seashore on the south coast.

In the afternoon we all meandered around aimlessly in what looked like a massive bomb crater, trying to interpret the geology at Seltangi, a task set by Dave McGarvie. Using self-taught skills of self-deception we were able to come up with a wide range of descriptions, explanations and answers, and with a charitable assessment of our efforts Dave was willing to fill in the missing 99% of the story (Fig. 38, *opposite*).

Seltangi, we were told, was an unusual crater-like feature that represented a ponding of magma, possibly where it has come into contact with the sea and ice. Inflation of the lava pool formed a blister (like a giant tumulus) until escape pathways were found

Figure 35 Eyjafjallajökull and its debris-filled glacial lagoon (© Geoff Downer 2011).

Figure 36 Eyjafjallajökul kettle hole ((© Bob Morley 2011).

for the lava, after which the structure subsided, leaving a crater floor of broken and upended-slabs that had been the crust over the lava lake. Numerous volcanic features within and around the structure helped both to clarify and confuse the issue.

It was with some regret that we then made our way to Reykjavík, as Seltangi was to be our last geological stop. We would shortly be back to civilisation, having seen over the past 11 days so many wonderful picturesque and geological sights that make up only a small part of Iceland's untamed scenery. We were all extremely grateful for Dave's excellent leadership, good humour and his willingness to share his vast knowledge of the island's culture and geology.

Figure 37 Hellisheidi Geothermal Power Station (© Shirley Blaco 2011).

Figure 38 Seltangi. Dave and Chris relaxing while we solve the geological puzzle (© Averil Leaver 2011).

Book review

Woodcock, Nigel H. and Strachan, Rob 2012 *Geological History of Britain and Ireland* (2nd edn). Oxford: Wiley-Blackwell (ISBN hardback 978-1-40519-382-5, paperback 978-1-40519-391-8; 456pp; hardback £75.00, paperback £37.50)

This is a book with an ambitious premise: to produce a full geological history of the British Isles, at a level that both student and professional geologists can engage with, in one single, comprehensive, yet succinct volume. Remarkably, it succeeds in all these goals. The first edition was published in 2000, and this second edition has been revised by the original authors, assisted by a number of new contributors.

The book is presented as a series of chapters, each of which could be considered as a stand-alone paper. Part 1 starts with a broad overview of historical geology: why it is important, how we date events, how we can track and model environmental changes, followed by a brief look at the place of the British Isles in a global context. The book, like the British Isles themselves, begins divided, with Parts 2 and 3 dealing with the land that started off north and south of the Iapetus Ocean, respectively; in Part 4 the Iapetus Ocean closes; Part 5 sees the resultant continent of Pangaea consolidate; and Part 6 discusses our intra-plate setting from the Permian onward. The layout of the book, although presented chronologically, makes it easy to dip in and out of the sections in which you are most interested without having to worry about the previous chapters, so if you are particularly interested in a particular period or region you can jump straight in instead of worrying that you have missed something.

As a fairly new level-two student, I found some of the book quite hard going. This was almost certainly because much of the subject matter is new to me, and I was regularly having to dig out my Earth Science dictionary to explain some of the terms used! The further you are into your studies and the more you know about geology, clearly the easier it will be to read. This is not to say that the book is beyond new undergraduates: a particular strength of the book is the excellent, detailed diagrams that complement and reinforce the text; and I found the sections that dealt with my 'home patch' particularly invaluable. Those studying at a higher level will also find extensive references handily listed at the end of each chapter in case more in-depth reading is required. Perhaps the only real disappointment is that there are few references to places where you can actually go and see the geology, and while a volume such as this could never be considered a field guide I think this would be of great benefit to undergraduates in particular.

Although my review copy was a paperback I have taken a look at a sample of the Kindle version and this seems to be well-produced with good formatting. The diagrams are obviously smaller but perfectly readable, and at £25 rather than £33 this may be worth considering for those who do not need the print version.

All-in-all if you are a UK-based geologist I think this is an essential book to have on your bookshelf. The British Isles have such a fascinating geological history and this book has enriched my knowledge considerably. I am sure it will be a book I come back to again and again over the years.

— Richard Blagden, current OU student in Earth Sciences, Cert Con Sci (Open)

Book Review

Ensom, Paul and Turnbull, Malcolm 2011 *Geology of the Jurassic Coast: The Isle of Purbeck — Weymouth to Studland*. Wareham: Coastal Publishing (ISBN 978-1-90770-100-9; paperback, 128pp; £9.95)

The Jurassic Coast was recognised as a World Heritage Site in December 2011 for its outstanding geology and geomorphology. The Jurassic Coast's geological history displays a near continuous sequence of 265Ma of rocks spanning the Triassic, Jurassic and Cretaceous periods. The rocks of these periods are easily visible in the steadily eroding cliffs, and there is a 95-mile walk from west to east that takes you gradually from the ancient to the recent.

This book is one of a series covering the Jurassic Coast, and actually covers more than just the Isle of Purbeck, for it runs from Weymouth to Old Harry Rocks, east of Swanage.

The first third of the book introduces the geology of the Isle of Purbeck, then describes the rocks of the Purbeck coast. These rocks are explained well in the text and are accompanied by many illustrations and diagrams; the highlights are the marked-up photographs, which are clear, informative and excellent (and are the best part of an excellent book). There are good explanations of the alternative theories for the formation of Lulworth Cove and of the Old Harry area, plus sections on the fossils and the fossil fuels of both areas.

The bulk of the rest of the book consists of a series of itineraries, working from west to east. These include two boat trips — illustrated with labelled photographs and detailed geological sections — all as viewed from the sea. Finally, there is an underwater three-dimensional map showing the "rocks, reefs and races" in relief, to show how the geology of the Jurassic Coast continues into the English Channel.

The book includes a glossary, an index, advice on visiting the area and a list of further reading.

This book sets the standard for geological field guides — the diagrams and annotated photographs are worth the price alone. When the book is carried and consulted in the field the geology becomes self-evident. Such a book can only both encourage more fieldtrips to the Jurassic Coast and a better understanding of the geology of the area.

— John Lonergan, OUGS London, BSc (Hons) Geosciences

Wessex OUGS excursion to Assynt, north-west Scotland — 5–11 May 2012

(Led by John Mendum, BGS, Edinburgh)

Mark Barrett, Alan Holiday, Alan Mack, Bob Nunn, Jane Schollick and Stuart Swales

Day 1 Scourie and Tarbet

by Stuart Swales

Owing to the uncertain weather forecast for Monday we decided to transpose the Day 1 and 2 itineraries. 'Day 2' was a trip to examine lithologies and structures of the Lewisian Gneiss Complex of the Foreland.

Driving north from Inchnadamph through pretty awful weather (rain and snow higher up), we were glad that it had cleared up somewhat by the time we reached our first stop on the coast near Scourie Mòr (Fig. 1).

Here John Mendum introduced us to fairly typical gneisses of the Assynt Terrane that crop out around the bay of Camas nam Buth — these were stripy grey felsic and subsidiary mafic gneisses, with a steeply dipping compositional layering/foliation. There are ultramafic pods found with the foliation swirling around them. The grey gneisses here are some of the oldest rocks in Europe; their tonalitic, granodioritic and doleritic igneous protoliths have given ages of up to 3030Ma (U-Pb zircon). During the Badcallian event at *c*. 2700Ma these rocks had then undergone considerable deformation and granulite facies metamorphism (involving much recrystallisation) to form the felsic and mafic gneisses. With not much subsequent deformation or retrogressive metamorphism seen here, these are allegedly some of the easiest bits of the Lewisian Gneiss Complex to interpret!

Slightly farther north-west, we encountered a knoll of ultramafic rocks, the lusher green grass here contrasting with the poor vegetation found covering the felsic gneisses. This ultramafic rock was compositionally variable, with the metamorphic mineralogy reflecting original igneous layering; there are large (*c.* 50mm) red-brown-purple magnesium-rich pyrope garnets in some layers (Fig. 2). The mineralogy suggests that the ultramafic protoliths may have crystallised at lower crustal depths. This knoll is a small exposure of a large, layered body of mafic and ultramafic gneisses whose relations with the felsic gneisses suggest that it predated their protoliths. The felsic gneisses seem to intrude the mafic and ultramafic ones, but Sm-Nd isotopic dates from the mafic-ultramafic and felsic gneisses contradict the field relationships. Perhaps their isotopic systems have been perturbed during the Badcallian gneiss-forming event. John told us that the ultramafic rocks here tend to be associated with metasedimentary rocks. This raises yet many questions about their origin — being Archaean, they could possibly even predate plate tectonics.

Figure 1 Our home for the week, Inchnadamph Lodge, with Conival (987m) behind it (© Alan Mack 2012).

Farther along, we found a microgranite sheet with a planar fabric and smeared-out quartz crystals, within the ultramafic gneisses; this rock could be traced north-west and seen to be discordant in a precipitous face above the bay, with metre-scale folds clearly visible in the gneisses here adjacent to the intrusive granitic sheet.

Crossing a storm beach and over a headland we proceeded toward Geodh' Eanruig, a north-west-trending vertical fracture zone prominently cut across the gneisses of the headland to the north. En route we found more mafic and ultramafic gneiss sheets with folded fabrics. Here we could easily see pale green clinopyroxene crystals, together with garnets, with darker hornblende after olivine. At Geodh' Eanruig the greasy-looking grey felsic gneisses were cut by a markedly discordant 200mm wide dark microcrystalline dyke with clean-cut margins trending *c.* 120°. This was the first Scourie dyke we saw on the excursion.

Mafic Scourie dykes had been intruded at lower crustal levels into the bulk of the Badcallian gneisses of the Foreland during the Palaeoproterozoic at *c.* 2400Ma. Here, the Scourie dykes had been metamorphosed under amphibolites-grade conditions, but remained undeformed, a feature typical of the Assynt Terrane. Near by, we came across a better three-dimensional exposure of a subvertical Scourie dyke, also cutting across the banding of the grey felsic gneiss, some 3–4m wide, trending *c.* 130°. We observed thin, near-parallel splays coming off this dyke, these being attributed to minor stress variations in the host rocks at the time of the emplacement of the Scourie dykes.

Figure 2 Garnet in the granulite facies (© L. Morley 2012).

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Figure 3 Scourie dyke (© L. Morley 2012).

Ultramafic pods in the gneiss here have lots of shiny black hornblendes after olivine/pyroxene, showing that hydrous fluids have been involved, with small shear zones along Scourie dyke margins. The grey gneisses are also cut by an undeformed pink pegmatitic granite sheet, with coarse graphic intergrowths of light grey quartz and pink alkali feldspar. This sheet is itself cut by a Scourie dyke, and thus its age is intermediate between the Badcallian gneiss-forming event (*c*. 2700Ma) and the intrusion of the Scourie dykes — this would bracket the emplacement of the granitic sheet as Inverian (*c*. 2500Ma) (Fig. 3).

Our return to the minibus detoured over the rocky knoll, where we found an odd exposure showing a pegmatite sheet that separated grey gneisses from a folded ultramafic pod — unusually, the tight folds apparent in the ultramafic pod appeared to continue into the pegmatite sheet. One explanation for this is that the pegmatitic material had slowly replaced the ultramafic minerals, but retained a 'ghost' of the original fold. Also en route we located outcrops of grey gneiss with abundant close to tight folds of the layering trending about west-north-west — a signature of ductile deformation during the (pre-Scourie-dyke) Inverian event.

John pointed out that while there was a lot of glacial till deposited on the Torridonian outcrop of the Foreland, there was much less so on the Lewisian outcrop, usually only small accumulations found in depressions. This could be down to changes in the hydrology at the base of the Quaternary ice sheet, the more porous Torridonian draining the wet-based ice sheet and causing more till to accumulate.

Lunch was taken on the hillside before heading farther north to the tiny village of Tarbet. Still within the Assynt Terrane, this locality is much closer to the Laxford Shear Zone (LSZ). The Inverian deformation increases markedly towards the LSZ, as documented by Sutton and Watson (1951) — a recent re-appraisal has proposed that the LSZ is a Laxfordian (*c*. 1750Ma) reactivation of an earlier Inverian terrane boundary.

On the rocky ridge behind the Shorehouse Restaurant at Tarbet we found flaggy to fissile brown weathering (due to pyrite?) and white streaky semipelites, with a strongly dipping fabric, together with very thin amphibolitic mafic layers. These are interpreted as metasediments of Archaean age, and are cut by Scourie dykes. Farther along, we found subhorizontal striae of glacial origin, with bigger grooves towards the top of the outcrop. In some gneisses we found up to 6mm red-brown almandine (iron-rich) garnets in darker grey layers also containing hornblende — possibly metavolcanic rocks?

Across the knobbly hillside we passed across a 'transition zone' of striped quartzo-feldspathic gneisses interleaved with amphibolitic mafic layers whose origin was a little problematic. Farther up the small stream valley was a spectacular outcrop showing thick amphibolitic mafic pods and layers with muscovite-biotite granite veins that themselves were tight to isoclinally folded. The mafic layers had a strongly defined fabric of plagioclase and hornblende near the contact with the felsic layers, suggesting that the latter were ductile, whereas the cores of the mafic layers, studded with tiny red garnets, remained 'dry'. The deformation (tight, upright folds) is of Inverian origin (Fig. 4, *opposite*).

Near by, on the west side of the valley, undeformed pegmatitic granite sheets cut sharply across earlier gneiss fabrics and folds — these were the first Palaeoproterozoic Laxfordian granitoid sheets and veins (*c*. 1600–1800Ma) we saw en route to the Big Scourie Dyke. On the north-west slope of Cnoc Poll an Turrabain there are textbook examples of Inverian folding of the Badcallian felsic and mafic gneisses. The folds are clearly seen in down-plunge section, and locally display somewhat discordant profiles, reflecting layer thickness and locally variable deformation intensity.

We came down to the coast near Poll an Turrabain, where there is a good viewpoint from which to see the Big Scourie Dyke. From here, it can be seen to clearly cut across the tens-of-metres scale Inverian structures, even though, strangely, the dyke terminates suddenly inland. From close up, things are more complicated! Strong Laxfordian shear zones deform the dyke internally, with good fabrics defined by stretched feldspars. The shear zones form an anastomosing three-dimensional network around much less stressed metre-scale pods with strong fabrics at their margins. Some views of the fabric, on vertical faces, show a prominent south-east-dipping lineation defined by feldspar and hornblende with crystals up to 20mm long (Fig. 5, *opposite*).

Our final stop was a few miles north of Laxford Bridge, at the Loch na Fiacall road cut — the famed 'Multi Coloured Rock Stop'. This locality lies north of the Laxford Front in the Rhiconich Terrane. Here Laxfordian reworking of earlier structures (viz. Badcallian gneisses and Scourie dykes) can be clearly seen in the clean rock face produced by the pre-split blast technique. It is a good place to work out the chronology of events for yourself, although there is an information board here to aid the befuddled of brain after a long day in the field!

Figure 4 (right) Inverian outcrop (© L. Morley 2012).

Figure 5 (below) Big Scourie Dyke (© L. Morley 2012).

Figure 6 Multi-coloured rock (© L. Morley 2012).

The sequence of events encompassed by the exposure could be briefly summarised as Igneous protoliths → Badcallian gneisses → Scourie dyke emplacement → Laxfordian pegmatite emplacement and deformation (Fig. 6).

Some of the mafic elements appear to be older than the Scourie dykes, as they show extensive alteration with thick selvedges of biotite and hornblende; they are also heavily veined and 'mashed up', atypical of Scourie dykes. The pegmatitic granite sheets and veins were emplaced at *c*. 12–15km depth during the Laxfordian event, and lie discordant to both the Scourie dykes and the earlier Badcallian gneissic layering. However, late-stage deformation after pegmatite emplacement apparently caused localised deformation of the grey gneisses marginal to the now more rigid pegmatitic veins. This generated open 'flanking folds' in the gneisses whose axial planes were orientated subparallel to the pegmatitic granite vein margins. One eagle-eyed mineralogist spotted chalcopyrite in the pegmatite.

And so back to Inchnadamph after an excellent day in the field!

Sutton, J. and Watson J. 1951 'The pre-Torridonian metamorphic history of the Loch Torridon and Scourie areas in the north-west Highlands, and its bearing on the chronological classification of the Lewisian'. *Quart J Geol Soc London* **106**, 241–307

Day 2 Loch Assynt Foreland Traverse and the Achmore Duplex

by Jane Schollick

On Monday May 7th we traversed the geological succession of the foreland, mainly on the roadside section along the north shore of Loch Assynt. The foreland is the unmoved region lying west of the thrust zones. It consists of the basement Lewisian Gneiss Complex (Archaean/Palaeoproterozoic) overlain by Torridon Group red sandstones (Precambrian), Ardvreck Group quartzites (Cambrian) and Durness Group dolostones (Cambrian/Ordovician).

We stopped first at a viewpoint across Loch Assynt ,where we saw the famous 'double unconformity' on the northern flanks of Beinn Gharbh. The unconformities were easy to spot as each rock type produces a distinctive type of scenery — the hummocky, rocky and grass-covered Lewisian gneisses in the foreground are overlain unconformably by a thick pile of horizontally bedded Torridonian sandstones, in turn overlain unconformably by beds of shiny pale grey Ardvreck quartz arenites and sills of Canisp Porphyry, dipping *c*. 15° east.

We started our traverse by Loch Assynt, where grey, strongly foliated Lewisian orthogneisses were intruded by a 9m wide

ultramafic Scourie dyke. The pyroxene-bearing felsic gneisses were derived mainly from a tonalitic protolith (*c*. 3000Ma) and contained lenses, pods and bands of mafic and ultramafic rocks. They had been metamorphosed at mid-crustal levels under granulite-facies conditions during the Badcallian event (2700–2500Ma ago). This is one of the oldest rocks in the UK!

We then examined the basal unconformity of the Torridon Group (location 2), where conglomerates and pebbly sandstones of the Diabaig Formation (basal Torridon Group) unconformably overlay highly weathered Lewisian gneisses. This unconformity represents a huge time gap of at least 1000Ma, as the Diabaig Formation was deposited *c*. 1000Ma ago and the youngest intrusions of the Scourie Dyke Suite are *c*. 2000Ma old.

The Diabaig Formation consists of two lithofacies, both infilling the uneven hummocky Lewisian gneiss surface. The first lithofacies is a poorly sorted, tabular-bedded, coarse sandstone with a muddy matrix and matrix-supported pebbles, composed of gneiss and quartz. The pebbles are rounded to sub-rounded, locally with a desert-varnish, and some wind-etched dreikanter pebbles are present. The second lithofacies is a better sorted, comprising coarse-grained sandstone with matrix-supported pebbles. Some beds have erosive bases and others have laminated fine-grain tops. The matrix in both facies is arkosic (contains significant feldspar) and contains some highly rounded millet-seed sand grains. The Diabaig Formation is a desert sandstone, with the bedforms showing evidence of deposition in alluvial fans and lacustrine environments.

Walking east along the roadside, up the succession, we could see the change to the thick trough of cross-bedded sandstones of the Applecross Formation (Torridonian Group) conformably overlying the Diabaig Formation. This is dominantly a coarse- to medium-grained sandstone containing angular grains of terracotta-coloured feldspar and quartz. Trough cross-bedding (with 100–200mm sets) is common throughout and was seen in three dimensions on top of a hill showing a current direction of *c*. 100°. These sediments were deposited in deep, perennially braided rivers. The Applecross Formation is up to 1,000m thick in the Assynt area and is responsible for distinctive shape of many of the Assynt hills, e.g. Suilven and Quinag (Fig. 7).

Next we walked across the basal Cambrian (Ardvreck) unconformity. This is not exposed here, but could be seen clearly near the summit of Spidean Coinich (Quinag), where pale grey weathering, white, well bedded arenites of the Eriboll Formation dip east at 12°–15° and overlie the horizontal Applecross Formation with marked angular inconformity.

The Ardvreck Group comprises the Eriboll Formation, made up of the Basal Quartzite Member and the Pipe Rock Member followed by the An t-Sròn Formation (Fucoid Beds Member and the Salterella Grit Member). The Ardvreck Group consists of clastic shallow marine sediments, deposited on the wide eastern margin of Laurentia.

The Basal Quartzite Member is a pale grey quartz arenite (i.e. not a true quartzite) showing planar tabular cross-bedding. The Pipe Rock Member is a distinctive, extensively burrowed, quartz arenite, which is found over a wide area in north-west Scotland and beyond. At the classic location of Skiag Bridge, the matrix has been stained red-purple by diagenetic iron oxide, but the Skolithos burrows (vertical 'pipes') remain white as a result of differential early diagenesis, so they are easily seen (Fig. 8).

The Fucoid Beds Member consists of rusty-brown dolomitic siltstones, containing many trace fossils, including Skolithos. The Salterella Grit Member is another cross-bedded quartz arenite containing fossils, among which are numerous tiny (*c*. 3mm long) white 'mollusc' shells known as Salterella. The Salterella Grit Member is overlain conformably by the lead-grey dolostones of the Ghrudaidh Formation (Durness Group), marking a major change from clastic- to carbonate-dominated sedimentation on the Laurentian margin. Burrows are still evident as mottled patches in the dolostones (Fig. 9, *overleaf*).

The Achmore Duplex is the lowest of a series of thrust zones at the western edge of the Moine Thrust Zone. At the base is the Sole Thrust and above it are a series of imbricate thrusts in the form of a duplex. Our aim was to count the number of 'horses' in this duplex by walking up the thrust zone from south-west to north-east, in the opposite direction to the thrusting. To do this, we walked uphill along a stream section west of Achmore Farm, noting the repeating successions of three rock types: rust-coloured Fucoid Beds siltstones (with primroses growing on them), grey Salterella Grit, and buff-grey, weathered Ghrudaidh Formation and paler grey Eilean Dubh Formation dolostones. Each of these successions comprises a 'horse' in the duplex: we counted at least seven of these 'horses', although the Ghrudaidh

Figure 7 Applecross, showing trough (© L. Morley 2012).

Figure 8 Pipe Rock Member (© L. Morley 2012).

Figure 9 Fucoid bed showing trails (© L. Morley 2012).

dolostone was frequently not exposed, as it had weathered away leaving hollows. At a waterfall we observed the thrust plane, dipping *c*. 30° north-east, where the Fucoid Beds are thrust over the Ghrudaidh Formation.

As we walked back down through the imbricated sedimentary sequence, we noted a sudden change in the dip of the bedding from *c*. 50° to 5°, close to the road at the bottom of a waterfall. This is where the floor thrust of the Achmore Duplex occurs — the Sole Thrust — that here marks the Caledonian orogenic front.

Day 3 The Traligill Valley

by Mark Barrett

We entered the Traligill Valley (from Norse, 'valley of the trolls') in light drizzle and under a lowering sky. We oriented ourselves with the surrounding terrain using our Assynt Excursion Guide. John Mendum pointed out how the river valley separates two different areas: to the north there are thick thrust sheets that predominately contain Lewisian Gneiss and Eriboll Sandstone Formation; to the south there are thinner thrust sheets that contain imbricates formed mainly from the Ardvreck and Durness groups.

A short way into the valley John drew our attention to some Cambrian Durness Group dolostone outcrops just north of the track. They exposed the light grey Eilean Dubh Formation that here was covered in clints and grikes. This karstic weathering normally occurs when a bare limestone surface is exposed to the effects of rainfall, but there was a brief but inconclusive debate on whether it had formed under a soil cover in this instance. A bedding dip here of 30° north-north-east is steeper than the 10°–15° of the same Cambrian sequence farther west. This difference is because the dolostones form a series of 'horses' (imbricates) that constitute part of a thrust duplex. As you travel upstream, you move farther into the duplex and consequently the bedding dips more steeply. The flora on the carbonate rocks was lush and varied but farther along the valley we noticed a dearth of flora on the hillside. Here, the limestone was overlain

Figure 10 The Traligill Thrust (© L. Morley 2012).

by glacial till and moraine, and hence calcium-poor soil. We saw both lateral and end moraine features, formed as the glacier retreated up the valley at the end of the last ice age, probably during the Younger Dryas, *c*. 12,000 years BP (Fig. 10, *opposite*).

Farther up the valley we noted the resurgence of the Traligill River, and could see a south-south-west-dipping planar surface of pale grey, weathering Eilean Dubh Formation dolostones sloping from the north into the dry riverbed. This surface represents a lateral ramp, which to the south dipped under a cliff of darker grey dolostones of the older Ghrudaidh Formation. The contact between the two formations is known as the Traligill Thrust. In the contact zone we found a rhyolitic intrusive sheet with iron pyrites. The small crystals therein indicated that the sticky felsic magma cooled quickly. The intrusion was not deformed and was intruded towards the end of the thrusting. Higher up the southern cliff face we could see another imbricate of Eilean Dubh limestone. Slightly farther upstream John Mendum also drew our attention to a mafic lamphrophyre (vogesite) intrusion. Following the thrust ramp along its strike we eventually came to the Lower Traligill Cave (sink), in which the Traligill Lateral Ramp is again exposed; and in Ghrudaidh Formation of its hanging wall a cave had formed. Within the cave and elsewhere along the river there is some unique flora (identified by Zoë Gardner) — particularly some types of ferns that flourish in this damp maritime climate.

In a dry side valley farther upstream John pointed out a whole series of imbricates and alternating ridges of Ghrudaidh and Eilean Dubh dolostones aligned roughly north-west within the river channel. Some of the beds within the 'horses' showed marked folding, indicating that thrust-related deformation was more ductile in these deeper structural levels. We followed the track uphill to the prominent cave entrances of Uamh an Tartair (Cave of the Roaring) and, higher up, Uamh an Uisge (Cave of the Water) (Fig. 11). The 'waterslide' entrance of Uamh an Tartair is along a thrust plane. These caves, through which the main Traligill drainage passes, developed in the hangingwall imbricates of Eilean Dubh dolostones that lie just above the main Traligill Thrust.

Above the caves looms the Cnoc nan Uamh klippe, an outlier of the Cnoc nan Uamh Thrust. The klippe contains a number of imbricates composed of units of the Lower Cambrian Ardvreck Group; in our ascent to the summit, we passed through the Pipe Rock Member, the Fucoid Beds and, finally, the Salterella Bed. The summit provides a panoramic view of the surrounding mountains and from here John gave us a geological interpretation of the landscape. To the north-east we could see where the Beinn an Fhurain Thrust emplaces the Basal Quartzite over Pipe Rock Member and where, higher up the mountain, the Beinn Uidhe Thrust emplaces Basal Quartzite over the top of Basal Quartzite. To the east we saw the brooding west face of Conival, which is mainly composed of the Ben More Nappe and includes folded Lewisian gneisses and overlying Torridon Group sandstones and conglomerates. To the south-east is Bealach Traligill, which marks a fault zone; and farther south the Breabag Dome.

Figure 11 Uamh an Tartair — 'cave of the roaring' (© L. Morley 2012).

This structure consists mainly of stacked imbricates of the Basal Quartzite and Pipe Rock members, with the overall duplex updomed by the underlying thrusts and horses to form an elongate broad dome *c.* 4km from north to south and 3km east to west. To the south we saw the dip slopes of Cnoc Eilid Mhathain, formed of the white weathering Eilean Dubh dolostones. Lastly, towards the south-west, another Torridonian sandstone outcrop on Beinn nan Cnaimhseag is interpreted as a klippe of the higher Ben More Thrust Sheet.

Stronchrubie Cliffs

After lunch, we drove a short distance south of Inchnadamph and viewed the Stronchrubie Cliffs from the roadside. John pointed out how the Ghrudaidh dolostones were topped by younger and paler Eilean Dubh dolostones, but that as you go higher up the cliff, the Eilean Dubh dips much more steeply east-south-east. These imbricates, which lie on the Sole Thrust, form a larger and complex duplex in the area. It is difficult to trace the exact location of the thrust itself within the cliffs. John also pointed out a pale weathering vogesite sill that ran through the dark grey Ghrudaidh Formation. This sill is part of a suite of undersaturated rocks that originated from melts deep within the mantle.

On the opposite side of the road, the flat valley is covered with glaciofluvial moraine terraces and drumlin-like features, left behind when a small infill glacier retreated from that part of the valley. While some people hold that this glacier originated during the Younger Dryas, cosmogenic age dates on some of the blocks suggest that it predated that event and is evidence for the prolonged decay of the Late Devensian ice sheets in this area. There is also debate on whether the glacier originated from the eastern flank of Canisp or from the western flanks of the Assynt mountains.

Travelling south toward Ullapool and Lairg we explored this small, disused quarry. The rocks therein formed part of the ultramafic to syenitic Loch Borralan Pluton. The main rocks in the quarry are melanite-augite-nepheline-syenites. These rocks contain white spots that are pseudomorphs after leucite, now altered to an amalgam of potassium feldspar, white mica and nepheline. This is the only locality in Scotland where you can find the sky-blue mineral cancrinite, a hydrated sodium aluminium silicate with carbonate. This used to be abundant in the blocks in the quarry, but over the years collectors have made it extremely difficult to find. One of our number, Andy Mack, found a very small sample of it.

The main rocks in the quarry are melanite-augite nepheline-syenites. These rocks contain white spots — pseudomorphs of leucite that have been altered to an amalgam of potassium feldspar, white mica and nightline.

Day 4 Durness

by Alan Holiday

The geology of northern Scotland is complicated. Offshore there are Mesozoic basins created by down-faulting, but onshore there is Lewisian rock (*c*. 2800–3000Ma), Moinian (*c*. 800–1000Ma) and rocks of Lower Palaeozoic age (*c*. 500Ma). The Moinian is down-faulted, also a surprise; but that is because of the effect of the Moine Thrust, which has pushed material up and to the west — so you can only see it at sea level if it is down-faulted.

At Durness the Durness Group is well exposed in Balnakeil Bay and it straddles the Cambrian Ordovician boundary. The Durness Group consists of five limestone formations, and the rocks young to the east. The oldest is the Eilean Dudh Formation or Black Isle Limestone; oddly, the limestone is a paler grey than the overlying Sailmhor Formation.

Durness Group

The Durness Group succession illustrates many features of carbonate sedimentology and it is a pity that it is so isolated. At the western end of Balnakeil Bay, before you reach the Kyle of Durness, is the island of Eilean Dubh (Black Isle), which is the type area for the lowest formation in the Durness Group. Working east over a distance of *c*. 1.5km you can see four of the five members of the Durness Group as the strata dip gently 10°–15°) east. The limestone is grey and fractured (not surprising, considering its age); it was deposited on tidal flats with fluctuating sea level and storm events. It accumulated in a peri-tidal environment, the sediment units bounded by flooding surfaces created by sea level rise or land level fall. Evidence suggests that these sediments were deposited in shallow water. Ripple marks occur as well as evidence of desiccation. This includes salt pseudomorphs, which suggest hypersaline conditions in shallow lagoons. There are also flake conglomerates where storm events ripped up the sea floor sediments. Within the sequence there is also evidence of millet seed sand grains (slightly frosted), suggesting that material had been blown in from a nearby arid continental environment. There is occasional evidence of evaporite sequences with ‘chicken-wire’ anhydrite (which has been replaced by quartz).

There is a variety of dolomitic limestone structures in the cliffs along the coast of Balnakeil Bay. The general term for the limestone in the Durness Group, which straddles the Cambrian Ordovician boundary, is microbialites. These limestones include stromatolites (with a laminated fabric), thrombolites (with a clotted fabric) and leiolites (structureless) (Fig. 12). Some beds showed more than one type of fabric. This bacteria-based organic activity occurred widely from the Precambrian (Palaeo-proterozoic), but the early Ordovician represents the last time when it is widely preserved in the geological record. (There are also examples in Dorset within the Purbeck sequence — e.g. the

Figure 12 Stromatolites, Balnakeil Bay (© L. Morley 2012).

Figure 13 Chert stromatolites, Balnakeil Bay (© L. Morley 2012).

Fossil Forest at Lulworth and on Portland at Kingbarrow and Weston.) Such bacteria-based organisms represent niche environments rather than a widespread development.

Within the Balnakeil section there is evidence of parasequences (in sequence stratigraphy terms) where sea level fluctuations resulted in sedimentary structures that can suggest shallowing and deepening of the water.

The limestone is dolomitised, which has affected its structure and possibly destroyed evidence of the original fauna, although evidence of trace fossils and some body fossils was found. Geomorphological evidence in the cliffs includes geos, stacks and arches.

There are similar stromatolite mounds on the Dorset coast in the Purbeck sequence, and in places chert has replaced the limestone with interesting dome shaped mounds, although the field notes describe these as being between the stromatolite mounds. In later limestone deposits chert and flint replaces limestone or fills burrows where acid conditions occur (silica is soluble in alkali/neutral conditions found in the sea but insoluble in acid conditions found in burrows). However, in the Durness case it is not apparent what the controlling factors were. The chert has a variety of forms, some persistent beds are 100–200mm thick while other masses are more organic in form. The amount of chert also fluctuates: some beds are rich in chert (e.g. parts of the Sailmhor Formation) and others are almost free of it (e.g. the lower part of the Eilean Dubh Formation) (Fig. 13).

The strata comprise obvious beds, their varying thickness providing evidence of fluctuating conditions. There were periods of erosion, as noted above. Gutter casts, where erosion occurs in channels and are infilled with later sediments, are another feature in some units. Owing to present-day erosion these features can be seen in three dimensions.

Some channels are straight, others sinuous. Evidence of compaction suggests that the sediment remained unlithified for some time. It is difficult to explain why the infill was more competent than the eroded material, however, when both sediment and infill look similar in present-day outcrops. Some units show ripple cross-bedding, again showing a mobile environment in the shallow water conditions. Occasional thin beds of siliclastc silts could be wind-blown, and one example seemed to drape over an erosion surface.

Sango Bay and Traigh Allt Chailgeag

Our second stop was Sango Bay. Here we saw evidence of the retrogressive metamorphism of the Lewisian material, caused by proximity to the Moine Thrust. Evidence for the alteration increases east to west across the beach. Initially, the rock is obviously Lewsian Gneiss, but within *c*. 100m it changes. The rock shows greenschist facies (chlorite rich) rather than the granulite or amphibolite facies seen elsewhere. Mylonitisation has occurred where the rock has been intensely sheered, creating a foliation emphasised by quartz veins that have been sweated out of the rock. Shearing has reduced the thickness to as little as 2% of the original. The highest unit of the Durness Limestone (Durine Formation) occurs at the western end of the bay, where the situation is complicated by thrusting.

The final locality of the day, Traigh Allt Chailgeag is just east of Sango Bay. Here there is impressive Lewisian Gneiss with excellent near-vertical banding. The banding is emphasised by mafic units (Scourie dyke material), which was reworked in the Laxfordian into boudinage structures in some cases (like a string of sausages) (Fig. 14).

Figure 14 Boudinage, Traigh Allt Chailgeag (© L. Morley 2012).

Day 5 — Unconformities, barchan dunes, glacial dropstones, the earliest life forms in the British Isles and an extra-terrestrial impact

by Alan Mack

Enard Bay

We parked in a lay-by [NC 0211 1241] on the string road between Stac Pollaidh and Achiltibuie. Visibility was good. In the distance, to the south were the Torridon peaks of Beinn Eighe, Liathach and Beinn Alligin; closer at hand Suilven and Canisp; to the far north Ben Stack, Arkle and Foinaven.

At Enard Bay we examined the stratigraphy and sedimentology of the late Mesoproterozoic Stoer Group and its unconformable relationships with the underlying Lewisian Gneiss Complex and overlying Neoproterozoic Torridon Group.

	Stratigraphy	
TORRIDON GROUP	Applecross Formation	
	Diabaig Formation: conglomerate	
	Diabaig Formation: Siltstone	994 ±48Ma[1]
STOER GROUP	Meall Dearg Formation	1177 ±5Ma[2]
	Bay of Stoer Formation: Pool a' Mhuilt Member	
	Bay of Stoer Formation: Stac Fada Member	
	Clachtoll Formation	
LEWISIAN GNEISS	Orthogneiss	>1500Ma

[1] Turnbull, M. J. M.,Whitehouse, M. J. and Moorbath, S. 1996 'New isotopic age determinations for the Torridonian, NW Scotland'. *J Geol Soc London* **153**, 955–64

[2] Parnell, J., Mark, D., Fallick, A.E., Boyce, A. and Thackrey, S. 2011 'The age of the Mesoproterozoic Stoer Group sedimentary and impact deposits, NW Scotland'. *J Geol Soc London* **168**, 349–58

Locality 1 [NC 0215 1335] — Achnahaird Bay: Meall Dearg Formation

Observation: The rocks here are reddish, mainly coarse grained, and composed of quartz and feldspar (arkosic feldspathic sandstone). Cross-bedding is evident, with 2–10m lenses of planar and trough cross-stratification, and a few gritty bands. The cross-beds dip gently west.

Process: These rocks are the result of fluvial deposition of locally derived material, some as high-energy deposition.

Environment: Riverine, possibly braided with alluvial fans; palaeocurrent to the west.

Locality 2 [NC 0222 1419] — Achnahaird Bay: Camas a' Bhuailte

Observation: Similar rock type to locality 1. Some grains are rounded and frosted. Mixed planar and cross-bedding. Lenticular bedding built up one on another; large scale (50m wide).

Process: Fluvial deposition of locally derived material, deposition energy changing; some wind-blown deposits.

Environment: A large river channel through an arid environment, with some wind-blown deposits; or — less convincing — a barchan dune in a desert.

We cut across the peninsula to Locality 3; we could see the basement Lewisian Gneiss Complex exposed in the distance to the east, recognisable from its hummocky appearance. Walking east on the westerly dipping bedding we were moving down in the Stoer Group from the Meall Dearg Formation to the Poll a' Mhuilt Member of the Bay of Stoer Formation.

Locality 3 [NC 0271 1464] — Camas a' Bhothain: Poll a' Mhuilt Member

Observation: A sea inlet facing north with its west cliff composed of similar rocks to localities 1 and 2, with a mudstone underneath it (*c.* 2m thick); contact with the mudstone is sharp. A gneiss ridge on the east side of the inlet, with an on-lap of brecciated rock composed of gneiss clasts (some over 1m wide) in a coarse-grained muddy sandstone matrix interbedded with the red mudstones. Striations trending north-west are evident on the surface of the gneiss on the ridge.

Process: Low energy, possibly lacustrine deposits changing suddenly to higher-energy fluvial deposition and associated slumping. There was glacial reworking, but it is unclear whether it occurred before, during or after deposition of Stoer Group.

Environment: A lake changing into a river with debris fans and slumping of brecciated sediments into hollows in the (pre-glaciated?) palaeo-landscape (later glaciated?).

We cut across the gneiss ridge to Locality 4.

Locality 4 [NC 0278 1463] — 'Salmon Bothy' bay — west: putative 'dropstones' in the Clachtoll Formation

Observation: Large blocks of Lewisian Gneiss embedded in underlying mudstones. Close inspection revealed that the block has depressed some of the mudstone layers while younger mudstone layers on-lap the block. The block looks similar to the gneiss in the ridge 10m above the point of emplacement (Fig. 15).

Process: These blocks are the subject of much heated debate. They are interpreted as either ice-rafted dropstones that fell into the sediment when floating ice melted (Davison and Hambrey 1996) or as syn-depositional mass flow deposits (Young 1999; Stewart 2002).

It is worth dwelling a little on this debate to illustrate how geologists can come to quite different interpretations based on the same observations.

Figure 15 A putative dropstone, 'Salmon Bothy' bay (© Alan Mack 2012).

Davison and Hambrey (1996) argue that early Stoer Group sediments were deposited under glacial conditions. Using the principle of consilience, they construct an argument to support this hypothesis based on evidence from several independent observations, the dropstones being one of them. Stewart (1997) gives a robust reply to counter each piece of evidence, maintaining that their observations can all be explained by fluvial, lacustrine and slumping depositional processes. On the dropstones Stewart comments: "These sediments appear to me to be proximal rather than distal, and evidently deposited in shallow water, so the inference that the dropstones fell from melting icebergs through water twice as deep as the basement relief (Davison and Hambrey 1996, fig. 8) is surprising. More probably the stones tumbled onto an ephemeral lakeshore from the basement ridge to the west, which now stands *c.* 8m above the dropstone horizon."

Davison and Hambrey (1997) found this comment, and others by Stewart, 'stimulating', and replied with detailed responses to Stewart's assertions, maintaining their position that early Stoer Group sediments were deposited under glacial conditions.

Our own observations suggested that the gneiss blocks were emplaced into the fine siltstones and mudstones at a particular stratigraphical horizon. They were then on-lapped by succeeding fine-grained sediments. This would be compatible with a single collapse or slumping event, or would require a unique time of iceberg melting.

The debate continues.

Environment: Fluvial or lacustrine (Stewart); or glacial deposition (Davison and Hambrey).

Figure 16 Zoë Gardner sitting on the oldest recorded life form in the British Isles (© Alan Mack 2012).

We watched seals at play in the bay as we stopped for lunch by the ruined bothy and continued to discuss the arguments about dropstones, before setting off to view the eastern side of the bay.

Locality 5 [NC 0286 1462] — 'Salmon Bothy' bay — east: basal conglomerate and algal limestone, Poll a' Mhuilt Member

Observation: Exposed basement gneiss forming palaeo-hills with infill of conglomerate containing gneiss clasts — similar to Locality 3. Some of these exposures include reddish laminated limestones (effervesce with dilute HCl) containing fossil algae . The limestone is overlain by the mudstones of the Poll a' Mhuilt Member. Some desiccation cracks are seen in the mudstones. The mudstones are overlain by a thick band of basal conglomerate, belonging to the Diabaig Formation, which marks the start of the younger Torridon Group (Fig. 16).

Process: A low-energy, shallow water environment, possibly lacustrine deposits changing suddenly to higher-energy fluvial deposition and associated slumping. In calmer conditions algae developed into stromatolite structures. The limestone was later stained red by overlying mudstones. Much later, there was high-energy fluvial deposition.

Environment: A lake changing to higher-energy river deposits. Calmer conditions in a lacustrine setting became suitable for life to thrive in shallow, fresh (perhaps brackish) waters in the form of algal mats that draped the palaeohighs and encompassed some of the gneiss blocks — at 1177Ma these are evidence of the oldest recorded life forms found in what are now the British Isles; *c.* 200Ma later, high-energy river sediments created the Torridon Group basal conglomerate.

Of course no hammering is allowed at this unusually significant locality, although some of the group found the odd eroded piece in the surrounding rock pools on the way to the next locality.

Locality 6 [NC 030 146] — Camas a' Bhothain: Diabaig Formation

Observation: Massive thick-bedded sandstones; conglomerates with sandstone clasts, some beds dipping steeply northwest, with brecciated sandstones intercalated between larger blocks; all sitting unconformably on the underlying Stoer Group.

Process: High-energy fluvial deposition and flash flooding, with later landslippings and/or slumping.

Environment: Flash flooding of arid highland regions onto palaeo-landscape with later slumping to generate the brecciated sediments.

Locality 7 [NC 0308 1456] — Camas a' Bhothain: Stac Fada Member

Observation: Bedded sandstones; fining-up sequences with accretionary lapilli at the top — the Stac Fada Member. Overlying conglomerate contains cobble-sized clasts containing accretionary lapilli — the Torridon Group Diabaig Formation.

Figure 17 Accretionary lapilli in the Stac Fada Member (the 'squashed peas' to the left of centre (© Alan Mack 2012).

Process: Mudflows related to ash clouds. Accretionary lapilli form when ash particles amalgamate in moist ash clouds and then fall to earth to create distinctive concentric pea-sized features in the rocks (Fig. 17). The ash cloud was previously considered to be volcanic in origin but Amor *et al.* (2008) propose an extra-terrestrial origin for the ash cloud in the form of ejecta from a meteorite impact. A long period of time followed — there is an unconformity — and then there was high-energy fluvial deposition and flash flooding.
Environment: Vulcanism or a meteorite impact followed 200Ma later by flash flooding of arid highland regions, creating brecciated sediments on the palaeo-landscape.

On our way east to the final locality we made a small detour to Rubh' a' Choin to see the gently north-north-west-dipping arkosic sandstones of the Applecross Formation. Some dewatering structures were present in this gritty coarse-grained sandstone.

Locality 8 [NC 0360 1467] — Headland north of Creag a' Choin Mhóir: Stoer and Torridon conglomerates

Observation: According to the guide we were expecting to observe two conglomerates overlying one another: the Stoer Group (Clachtoll Formation) below and the Torridon Group (Diabaig Formation) above — separated by a sharp unconformity.

However, we could find no change in lithology across the obvious east-dipping feature that looked superficially like an unconformity or a sinuous discontinuity of some sort. Stewart (2002) argues that the Diabaig conglomerate has clasts of sandstones, whereas the Stoer one does not. This distinction cannot be made easily in the field at this locality (although it is easily seen farther south at Rubha Dunan — something to leave for another day) (Fig. 18).

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Epilogue

by Alan Mack

As we headed for the minibus we discussed among the group the events of the fieldtrip. Some thought that the earlier days had been more interesting. I thought that the final day, in which we had seen two unconformities (representing long periods of time), possibly barchan dunes, glacial (or lacustrine) dropstones, evidence of the earliest life forms in the British Isles, and evidence of extra-terrestrial impact were of more than enough interest for me. All this just underlines what a fantastic region this is for seeing outstanding geology. If you get the chance to go then put everything else to one side and sign up for the next trip to Assynt.

Special thanks to John Mendum (leader), who also compiled an excellent excursion guide for the week; to Mike Barrett (organiser) and to all the OUGS group from Wessex and farther afield for a fantastic field week!

Figure 18 Searching for the Stoer Group (Clachtoll Formation) — Torridon Group (Diabaig Formation) unconformity.) (© Alan Mack 2012).

Archaeological addendum

by Bob Nunn

Some of the group visited two Neolithic tombs near the Inchnadamph Lodge. Beginning in the 4th millennia BC, along the Atlantic seaways of Brittany, Britain and Ireland, such tombs follow a burial tradition spanning almost 1,000 years; a tradition especially well represented in the north of Scotland, in the Hebrides and in Orkney.

Both tombs are damaged. The Chambered Tomb at NC 23420 25520, by the road cutting, and the possibly later, second tomb at NC 26123 21774 — described by the OS as a 'cairn' — has been robbed of stone. However, from their positioning it can easily be seen how they would have dominated the landscape, probably acting both as territorial markers and as a 'home' for ancestors overseeing day-to-day life. The route to the sea and north passes the first tomb and the 'cairn' controls access to Gleann Dubh, its caves and pasture.

In use, the tombs would have held the remains of many individuals, possibly those of a 'chiefly' lineage. In some such tombs complete, crouched skeletons are found, but more often there are mixtures of bones. It is thought that perhaps after excarnation had defleshed the corpse, skulls and long bones were sometimes selected for special treatment (Figs 19 and 20).

Figure 19 (above) Collapsed chambered tomb (© Bob Nunn 2012).

Figure 20 (below) Gleann Dubh 'Cairn' (© Bob Nunn 2012).

Book Review

Bergslien, Elisa 2012 *An Introduction to Forensic Geoscience*. Oxford: Wiley-Blackwell (ISBN 978-1-40516-054-4; paperback, xxxii + 498pp; £37.50)

Being an avid watcher of the TV series CSI, NCIS and Body of Proof, this book grabbed my attention as soon as I received it and I decided that I would be the one to review it. And I have enjoyed it. Elisa Bergslien PhD is Associate Professor at Buffalo State College, New York, and as such the book has been written very much with American students in mind. However, she is also aware of a wider audience to which it would appeal and many of the examples she quotes come from Britain.

It is a hefty tome — more than 500 pages. The book is Quarto size (smaller than A4 [in fact, Quarto = 'American' paper size of 8.25in × 11.75in — Ed.]). Being printed on shiny paper, it can be a bit difficult to read in the 'wrong' light. There are three 'Indices': Lists of Tables and Figures, Colour Plates and Cases, in addition to a thorough List of Contents. These are at the start and there is a General Index at the back. Nevertheless, it is a fascinating read.

Elisa starts with a brief history of forensic science and crime scene basics. Then a further 10 chapters follow on minerals, rocks, maps, gemstones, sand, soil and fossils. She has included a chapter on the Geology of Art — helpful in proving forgeries. The final two chapters cover 'Forensic Anthropology' and Archeology', and 'Environmental Forensics', describing how to track pollution to its source.

Each chapter begins by explaining why a forensic geoscientist might need this information (e.g. art fraud, The Great Diamond Hoax), followed by the 'technicals' of the subject (e.g. what hardness in a mineral means, how to measure it, how it helps identification). Each chapter ends with a summary, a suggested further reading list and a list of references.

While I have no intention of becoming a forensic geoscientist, I found this book to be an excellent revision for basic geology — Elisa has aimed it not just at geology students but for any forensic scientist. Sometimes, I had to think carefully about what she meant when 'Americanisms' are used — but that's good from my point of view. In that respect it has given me an insight into how things are viewed 'over the Pond'. I probably wouldn't have gone out to buy this book but I am really glad I've read it; some of the cases Elisa has quoted are fascinating!

— Jane A Michael BSc (Hons) Open (Natural Sciences with Earth Sciences), OUGS Review Officer/Archivist

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Book review

McGowan, A. J. and Smith, A. B. (eds) 2011 *Comparing the Geological and Fossil Records: Implications for Biodiversity Studies*. London: The Geological Society, Special Publication 358 (ISBN 978-1-86239-336-3; hardback (laminated cover), vi + 247pp; £90)

McGowan (University of Glasgow) and Smith (Natural History Museum, London) have edited 14 papers under the unifying theme of the rock and fossil records and what comparison between these two datasets can tell us about observed palaeobiodiversity.

By way of preface or introduction they have authored the first paper, entitled 'The ties linking rock and fossil records and why they are important for palaeobiodiversity studies'. The key points in this introductory paper are: that there is a correlation between the quality of the geological record and the diversity of the fossil record; that what creates that correlation is not completely understood and is differently interpreted by geologists and palaeontologists; that the correlations vary across different environments, and that the consistency of this variation needs to be determined; and that palaeontologists should investigate this problem by first addressing the 'geological bias' in their data.

The 13 papers that follow, not surprisingly, range across geological time periods and continents, and all but two of them are written by two or more authors at universities and research groups throughout the world. Thus a range of geological and palaeontological expertise has been marshalled to explore these issues. Some papers are more thematic, addressing the issue of criteria for preservation of fossils in the geological/environmental matrix; others are specific to a region, a time period, a fossil type or a fossiliferous environment. All of these papers are technical and written primarily for specialists.

Nevertheless, any geologist, of whatever 'level' in their studies, will be able to derive much from these papers, even if reading them is a challenge owing to their specialist nature.

L. Cherns and V. P. Wright address the issue of skeletal mineralogy in marine invertebrates, seawater chemistry and the effects of this chemistry on fossil lagerstätten. They call one aspect of any variation the 'missing molluscs' effect. Comparing 'aragonite' and 'calcite' seas, they find similar skeletal mineralogical loss in both environments. Concentrating on mollusc fossils, particularly up to Triassic times, they conclude that fossil size matters more than sea chemistry.

B. Hannisdal has written a highly statistical paper, suggesting a non-parametric method for resolving discrepancies in 'rock-record sampling bias' and 'common response to external forcing'. This involves quantifying the relative importance of incomplete data and proxy data, then using the non-parametric method to detect/distinguish 'causal interactions' from 'incomplete and irregular time series' (i.e. spurious relationships). The methodology is applied to Phanerozoic oxygen, carbon and sulphur isotope records in marine carbonates, identifying 'complex interactions between climatic changes and carbon and sulphur cycling. Hannisdal's paper is definitely not for the statistically timid!

A. O'Connor, C. Moncrieff and M. A. Wills contribute a paper, also heavily statistical, on variation in stratigraphic congruence in Phanerozoic higher taxa. The hypothesis or premise is that cladograms show stratigraphic congruence, and that strong or weak congruence has heretofore been interpreted as a measure of the 'accuracy of cladograms' and 'fidelity of the [fossil] record'. They suggest that poor congruence may reflect an inaccurate cladogram, a misleading fossil record, or both. They demonstrate that most 'congruence indices' are in fact biased by non-uniformly distributed parameters both across taxa and through time. They examined 650 published cladograms to test these ideas, applying a range of multivariate models. Their methods revealed that 74.5% of the variance in stratigraphic congruence in their sample is explained by factors other than by congruence *per se*. They conclude that stratigraphic congruence is strongly influenced by

[continued on page 44]

The landslide at Banc Dolhelfa, near Llangurig, Powys

Colin Humphrey

Summary

The site, *c*. 5km south of Llangurig, is one of the largest bedrock slides in Mid Wales until now undocumented, though noted on the British Geological Survey Rhayader Sheet. It is 650m long downslope (east-north-east to west-south-west) and 150m to 200m wide. The line of the A470 road along the eastern side of the River Wye bulges conspicuously to maintain contour around the lower part of the slide mass. The landslide can be glimpsed driving north on the A470, but the whole feature is plainly visible from a minor road running along the western side of the Wye (Fig. 1).

Most landslides in the area are in superficial deposits, but here the Silurian bedrock has failed. This is a complex landslide. Slide mass displacement has left the south flank of the depletion zone with a bedding-parallel slope inclined towards the slide axis, and the north flank with a steep escarpment. Planar failure of thin-bedded mudstone occurred on the south side, and a pronounced subvertical jointing then caused blocks to be detached from the north side and from the head of the landslide. Depletion zone geometry is clearly translational in form, not rotational, with combined bedding-related multi-planar failure and jointing-related block failure. The site of the original failure is close to the visible fault parallel to the upper south edge. The slide may have been triggered by movement on this fault. Subsequently there has been some further collapse and retreat of parts of the northern escarpment. Screes show that rock displacement has continued from the upper part of the slide in recent times.

The slide mass itself is now inactive but movement was reported in 1850. The complexity of the site is of interest, particularly now that more attention is being paid to landsliding in Britain.

No rights of way exist over or near the landslide but experienced visitors can obtain permission to visit the site. Many parts are dangerously steep and there is also a risk of rock fall.

Geological history and regional structure

Palaeozoic

All the rocks of this part of Mid Wales were deposited in an ancient sea, the Welsh Basin, between 520Ma and 410Ma. The basin was situated near the northern edge of a microcontinent, Avalonia, itself on the southern margin of the Iapetus Ocean (e.g. Brenchley *et al.* 2006). During the Ordovician and Silurian periods sediment flowed into the subsiding basin from the Midland Platform to the east and from Pretannia to the south.

Cross-stratification in rocks of the north escarpment (most of which have moved) indicate palaeocurrents from between south-south-west and south-south-east. Tectonic forces slowly closed

Figure 1 The whole landslide site viewed eastward on a minor road west of the Wye Valley from SN 916 746, 1km from the head of the slide. The landslide toe extends to the river, the line of which is marked by the nearest trees.

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the Iapetus Ocean and Avalonia drifted from the southern to the northern hemisphere where, *c*. 415 to 390Ma, in the Early Devonian, it collided with the continent of Laurentia in the Caledonian event. The final stage of this collision, the Acadian Orogeny, led to uplift, and folded and cleaved the basin rocks during the early to middle Devonian. Banc Dolhelfa lies on the western flank of a moderately folded district, the north-north-east-trending Tylwych Anticline, a region of second-order folding between the first-order structures of the Central Wales Lineament 5km to the north-west and the Tywi Lineament 10km to the south-east (e.g. Davies *et al.* 1997). A steeply dipping east-north-east-striking fault lies very close to the southern flank of the landslide, passing beneath it at depth.

The rocks exposed at Banc Dolhelfa were laid down *c*. 430Ma. They belong to the Glanyrafon Formation of the Telychian Stage of the Llandovery Epoch, in the early part of the Silurian Period. At this location the BGS map cross-section (BGS 1993) indicates that the formation is now some 300m thick, much more having been eroded away. It consists of sandstone–mudstone couplets, the sandstones typically 10–30mm thick on the south side, which were deposited by turbidity currents from the southern shelf of the Welsh Basin. Such flows probably occurred on average every hundred years or so. The turbidites are Bouma Tcd types, with convolute bedding and cross-stratification in the sandstones (discussed in Davies *et al.* 1997). There are trace fossil burrows on the base of some sandstones.

Above the Glanyrafon mudstones in the stratigraphic sequence is the Pysgotwr Grits Formation, several bands of which are exposed in the vicinity by folding. One such band extends southward to terminate immediately adjacent to the upper northern escarpment of the Banc Dolhelfa site. These Grits are described on the BGS Rhayader Sheet as high matrix sandstone, locally pebbly.

Quaternary

During the last 2Ma, the Pleistocene Period, most of Wales was periodically covered by ice sheets and glaciers (e.g. Lewis and Richards 2005). The glaciers advanced during the glacial periods and retreated and melted during the interglacial periods. The ice sheets of the last glacial period, the Devensian, began *c*. 122,000 years ago and covered this area of Wales soon after, reaching a peak *c*. 25,000 years ago and finally wasting *c*. 15,000 years ago. The shape of the valley on this part of the Wye indicates the action of glaciers flowing from Plynlimon. The flat tops of the hills in the area correspond to the Welsh peneplain. Glacial push may have slightly oversteepened the bedrock of the lower valley side, leaving it less stable when the confining pressure of ice was removed; and glacial retreat left the lower valley sides mantled with till. After the river Wye was re-established, an erosive torrent of meltwater and gravel debris may have further undercut the oversteepened valley side, creating yet another source of instability. Freeze–thaw has led to continued erosion of exposed rock, as the extensive screes show.

Landslide dimensions and classification

The landslide is roughly rectangular in plan, *c*. 150m wide at the head and through the depletion zone, up to 200m wide at the widest point of the accumulation zone, and 650m long from the head to the tip of the slide mass at the bank of the Wye (Fig. 2). The slide displaced about one-third of a million cubic metres of rock, some of which remains within the depletion zone. The crown is at 465m elevation, and the landslide toe on the Wye is at 260m, giving the landslide a total height of 205m and an average gradient of 18º. The A470 road crosses the slide mass at 295m elevation, bulging conspicuously to the west to maintain its contour.

The British Geological Survey adopts the commonly accepted international classification for landslides, and presents on its

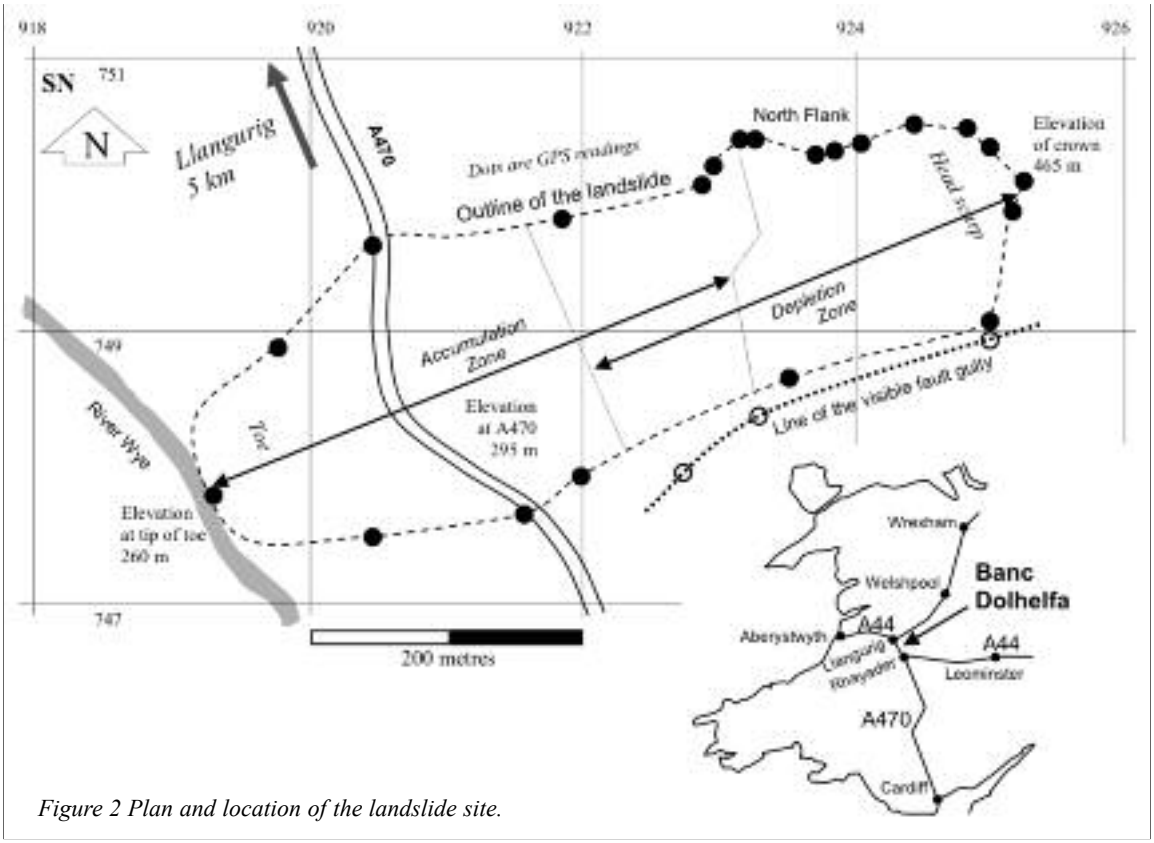

Figure 2 Plan and location of the landslide site.

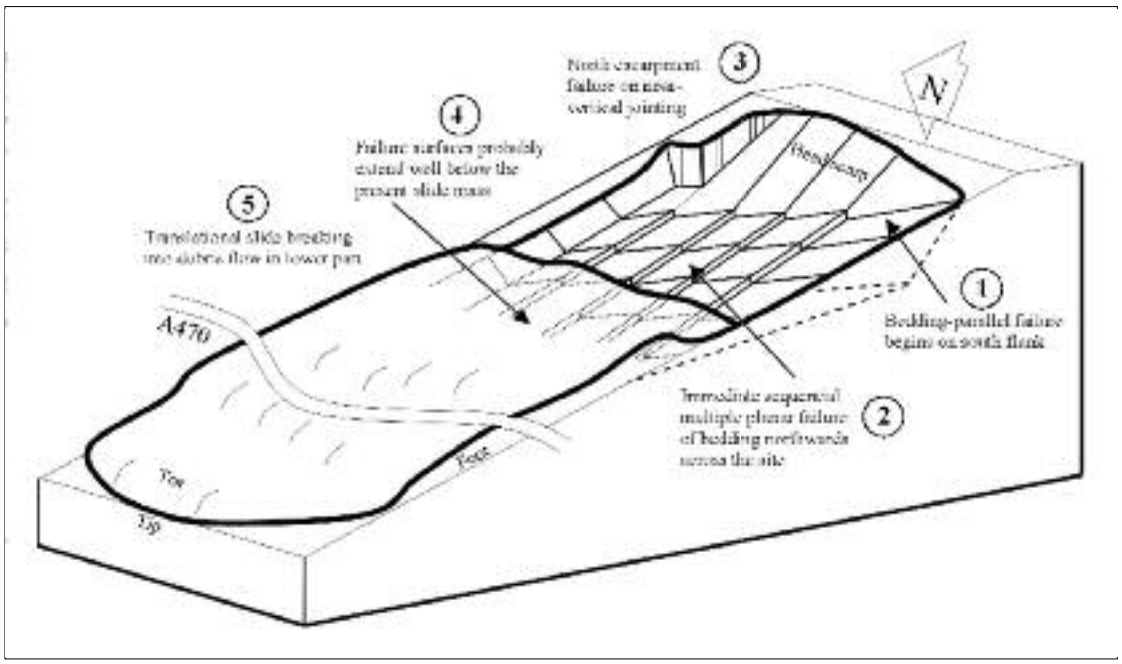

Figure 3 The translational nature of the slide, beginning with bedding-parallel failure and ending with joint-related failure.

website an illustration from Jones and Lee (1994), which is based on the seminal work of Varnes. Using this classification, Banc Dolhelfa, [grid ref. SN 9253 7502 to SN 9193 7478] can be described as a complex landslide, predominantly translational, with evidence of planar rock and debris sliding on the south side of the depletion zone and block failure by falling on the north side. Failed rock on both sides broke into a debris flow in the accumulation zone. Evidently the south side slid on bedding planes partly concordant with the topographic slope, and the back and north side failed on subvertical joints, leaving a steep escarpment (Fig. 3). Subsequent progressive failure and retreat of the escarpment has occurred by slumping and rock fall.

There is no evidence of rotational failure, nor is there hydrogeological evidence in the form of water accumulation or flow, to suggest that elevated pore pressures might be implicated in the failure. The very close proximity of a large fault parallel to the depletion zone suggests the initial failure may have been triggered by an earthquake caused by post-glacial movement on this fault, which is interpreted on the BGS sheet as passing below the area of bedding-parallel failure at depth. Many thin volcanic bentonite horizons were deposited in the Llandovery Epoch but no bentonite layers have been observed at this failure, and even if such a lubricant were to have assisted the initial failure, this is a multi-planar slide, and bentonite is unlikely to have existed at all the failure planes. There are no other recorded landslides along this part of the River Wye.

The main slide is evidently post-Devensian, but without finding buried wood it cannot be dated. However, the steepness of many debris slopes and hummocks, and the very thin soil cover, suggest recent origin, perhaps within the last 1,000 years. Apart from slowly continuing slumps and falls from the escarpment, and slow freeze–thaw degradation of exposed bedding planes, the landslide is now inactive. Although there has been some separation of small areas of debris since the initial event, no recent cracks are visible in the slide mass. The many stunted trees show no evidence of tilting due to creep. The road has been on its present alignment since the 18th century, and has been macadamed since the 1900s with no known serious maintenance problems (Mid Wales Trunk Road Agency, pers. comm. 2009). The BGS memoir briefly mentions a report of movement in 1850. The lead author of the memoir (J. A. Davies, pers. comm. 2009) recalls a newspaper report, held in the BGS archives, referring to a fall blocking the carriageway.

South flank and planar sliding

The hillside immediately adjacent to the south edge of the landslide has a slope of 28°. The south flank of the depletion zone is mostly vegetated on a very thin soil, but in places it exposes a single horizon of thin mudrock, dipping 43° west-north-west, towards the central axis of the landslide zone (Figs 4 and 5, *overleaf*), and striking with the regional trend south-south-west (206°).

The topographic slope of the hill intersects this bedding plane on a straight line west-south-west (252°) down the ridge that forms the south edge of the depletion zone. Thus, while the bedding dip is 43° towards the slide axis, the inward slope of the present south flank is only about 25° towards the slide axis.

Figure 4 (previous page) Standing on the intact crown, looking south-west down the upper part of the south flank, where bedding-parallel separation of the slide mass first occurred on a thin mudstone bed.

Figure 5 (right) Looking west down the vegetated slope of the bedding-parallel south flank in the lower part of the depletion zone, onto the hummocky accumulation zone above the road. The small ridges to the right are vegetated minor escarpments associated with further bedding-parallel slide planes.

The depletion zone lies below the level of the adjacent hillside for more than 200m downslope, after which the accumulated debris lies above the adjacent hillside.

Two prominent ridges lie on the floor of the depletion zone parallel to the south flank, with the same dip and strike as the south flank, and the same suggestion of bedding-parallel failure. The sandstone beds exposed on these ridges are up to 150mm thick (Fig. 6). These resistant beds define the floor of the depletion zone. Cleavage dips slightly more steeply than bedding, thereby assisting downslope movements. Bedding-parallel sliding probably initiated at the south flank, with failure quickly occurring on successive beds until it reached the north flank. The prominent ridges record the places where failure switched from one bed to another.

North flank escarpment and joint failure

The north flank of the depletion zone is an escarpment, once up to 25m high in places, and now with talus at its base. Most of the face of the escarpment is slumped on the jointing, leaving obvious dislocation of strata across the joints and a variation in apparent dip. In one place a 60m length of the face of the escarpment has slumped, collapsing more at the downhill end than at the uphill end, as can be seen from the angle of bedding (Fig. 7, *opposite*).

The shape of the escarpment slope debris at the head of the landslide on the northern side suggests that the initial slide extended all the way across the head of the site. Lower down the slide the jointing on the north escarpment has formed numerous columns *c*. 2m wide. Many joints have opened into fissures, leaving columns detached from the bedrock behind them (Fig. 8, *opposite*).

Figure 6 Standing on a bedding-parallel failure in the middle of the upper depletion zone, looking west at a slight escarpment, the other side of which was another bedding-parallel slide plane. Successive horizons of slightly thicker fine sandstones in the Glanyrafon Formation rock probably helped to form a floor to the depletion zone.

Figure 7 Looking across the scree that covers the head of the slide, at the steep upper north escarpment, with a 60m long vertically collapsed block in the centre view.

Figure 8 The northern escarpment in the lower depletion zone. The face consists of numerous back-tilted columns subvertically fractured on the jointing. At the far left (west) of the photograph the jointing ceases and the escarpment becomes less steep.

Figure 9 Looking up the slide at the whole of the north escarpment. The steep and jointed character changes abruptly, perhaps due to the close proximity of the more competent Pysgotwr Grits Formation. The ridges are associated with successive bedding plane failure across the slide from right to left.

The joints were probably initiated at depth in the rocks during various periods of erosion following uplift, from the Upper Palaeozoic to the Tertiary. These joints were then opened by freeze–thaw at the surface during the Quaternary, following glacial unloading. This has caused extensive subvertical jointing which strikes 310° and dips south-west at 80º, thus leaning away from the south-westerly direction of travel of the original slide mass from the north flank. Therefore the remaining escarpment tends not to topple, although talus indicates some collapse since the landslide.

The nature of the northern escarpment changes suddenly near the lower (western) end of the depletion zone (Fig. 9). It becomes less steep and is not as jointed as the mudstones of the steeper escarpment. The depletion zone itself occurs entirely in thin-bedded Glanyrafon Formation, which is mostly mudstone, with some thin sandstone beds forming the depletion floor. However, the Pysgotwr Grits Formation lies just north of the north escarpment of the landslide. In some localities north of Banc Dolhelfa these Grits are exposed in small quarries as massively bedded sandstone.

A change in lithology begins to appear on the lower northern edge of the Banc Dolhelfa depletion zone. Here, *in situ* sandstone beds around 200mm thick can be seen, with sandstone making up more of the rock than the alternating mudstone. In the scree this sandstone is coarse (grains >1mm) and may signal the beginning of the change of lithology into the overlying Grits. This rock, being more competent than the Glanyrafon mudstone, may have been a factor in preventing further northward spread of sliding. An area of small-scale tectonic folds of bedding occurs near by (Fig. 10).

Accumulation zone

Most of the accumulation zone is fairly uniform in distribution, with a little more debris on the north side where the hillside is slightly higher. This indicates a nearly full width translational slide. Thereafter the northern side of the depletion zone received additional debris, evidently from the middle part of the northern escarpment, either during the same event or later, and this has been retained within the excavated depletion zone. There are steep slopes and pronounced hummocks in the accumulation zone (Fig. 11, *opposite*).

Some of these hummocks expose large blocks of thin-bedded rock, which have not tumbled to destruction but have survived unbroken during translational sliding, and sometimes preserve a bedding geometry not much different to *in situ* rock higher up the slide. Where these blocks came to rest a mound formed around them (Fig. 12, *opposite*). Some large blocks tumbled to the bottom of the slide. One is visible in the river bank just above the waterline, with its bedding now vertical, in striking contrast to the gentler attitude of bedding in the *in situ* rocks just below it on the river bed (Fig. 13, *opposite*). Although the landslide clearly extended to the river, there is no evidence of river alignment being

Figure 10 Small-scale tectonic folds of bedding near the bottom of the north escarpment.

Figure 11 (right) Hummocky accumulation zone, looking west down the slide.

Figure 12 (centre) Large blocks helped to stabilise the slide and create mounds in the accumulation zone. They display bedding orientation, which suggests that they slid rather than tumbled in descent.

Figure 13 (bottom) A large tumbled block from the upper landslide resting at the tip of the landslide toe on the east bank of the River Wye.

significantly affected by the debris. The bank nearest the landslide contains slide debris, but the opposite bank does not, so it may be inferred that the landslide toe extended to the present river and little or no farther. The river may not therefore have carried away much of the landslide.

The fault

The geological map (BGS, 1993) shows a fault running down the slope of Banc Dolhelfa, striking east-north-east, one of the major faults in this region and thought to have originated during a late stage of the Acadian Orogeny (Davies *et al.* 1997, 195). This structure, more than 20km long, is represented on the western side of the Wye Valley by a huge cleft in the hills (Fig. 14, *overleaf*). It is not visible crossing the till of the valley floor, but is represented on the eastern side by a deep fluvial gulley on the upper part of Banc Dolhelfa (*see* Fig. 1). The southern edge of the landslide depletion zone clearly lies parallel to this fault gulley, and 15m from it, the two features continuing together for 200m downslope. Thereafter the line of the fault on the geological map curves south away from the landslide and the gulley disappears before the fault reaches the road. On the map cross-section the fault is interpreted to be the southern edge of a horst, with an up-throw between 200m and 300m and a steep dip to the north-north-west, which takes it below the landslide at depth.

Many rock fragments in the debris near the south side slope show deeply slickensided polished surfaces, some of which are mineralised with quartz; they are interpreted to be derived from the fault zone and to record movements within the zone (Fig. 15, *overleaf*). They are not seen on the north side distant from the fault. The presence of polished planes along the fault zone may have increased rock instability in this area. It is possible that an earthquake, caused by post-glacial reactivation of this fault, triggered the landslide. From the western side of the Wye another two fluvial gullies, representing minor faults, are clearly visible 500m to the north; though without associated landsliding.

Accessibility and safety

The site includes a ravine in the hillside rising from the A470 trunk road connecting Llangurig with Rhayader. The whole of the landslide site east and west of the A470 is private land used for rough grazing. Prior permission to visit the site must be sought

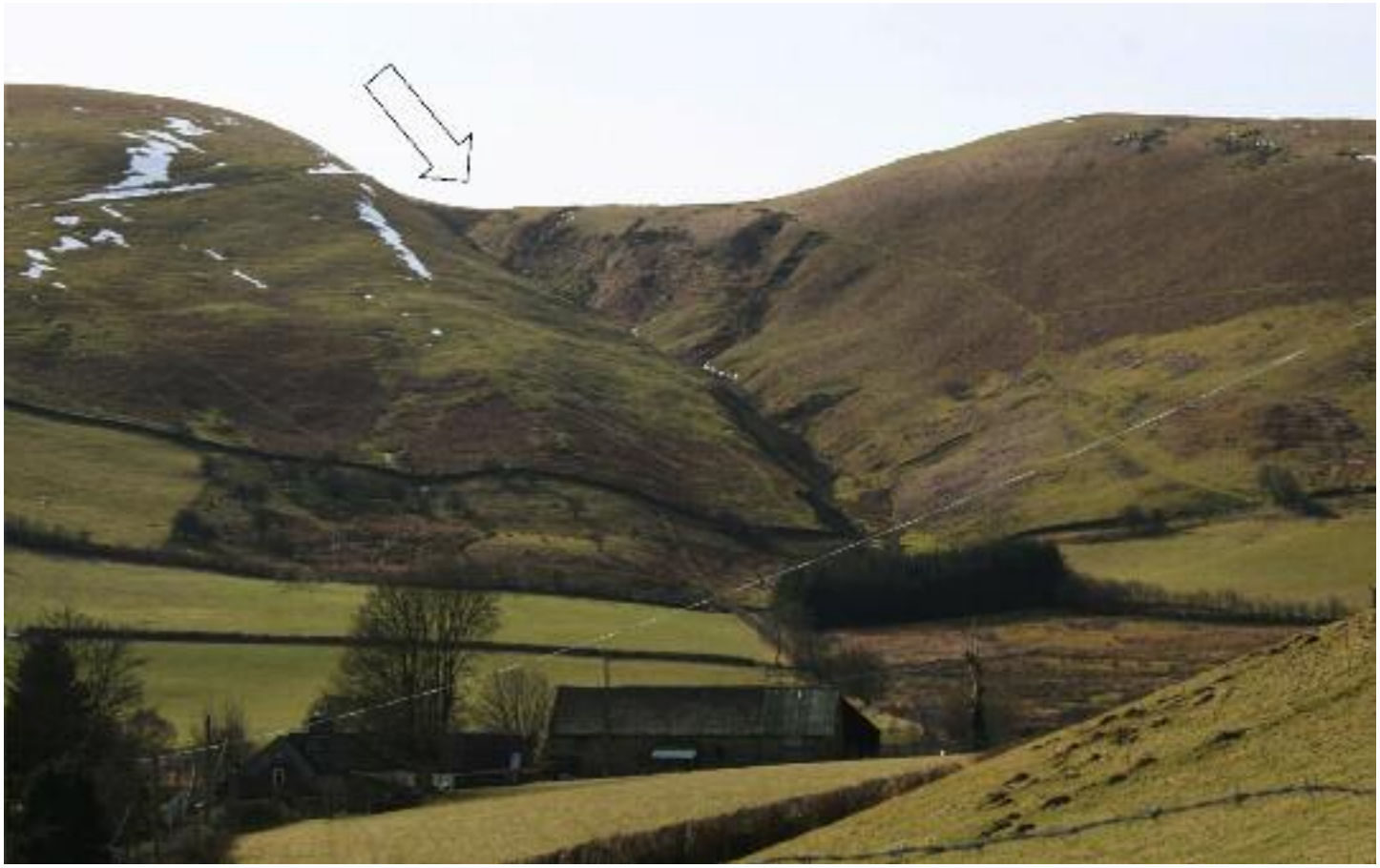

Figure 14 Looking west from the road at 1.4km distance, the fault beside the landslide reappears as a huge cleft on the opposite side of the Wye Valley.

from a farm [SN 913 763] several kilometres from the site. The RIGS is best accessed from below, through a fence gate from a small lay-by [SN 922 748] on the A470 near the southern edge of the slide mass, which can accommodate a coach. The landslide can also be viewed from minor roads across the Wye Valley at distances of *c.* 1km (*see* Fig. 1). One viewpoint is on the Severn Way trail, running along the western side of the valley. Off-road parking is available [SN 914 748], from where good views of the entire landslip and fault gulley are also obtained.

The site is dangerous and should only be entered by people with experience of such terrain, and not during wet or windy weather. The head and south flank are very steep and the north flank is perilously precipitous. The depletion zone is subject to rock fall.

Acknowledgements

The author is grateful to Tony Thorp, Bill Bagley and other colleagues in Mid Wales Geology Club for help with the survey, and to Professor Bill Fitches, chairman of Central Wales RIGS for advice during the work and in drafting the report. The site is recorded as Welsh RIGS number 578 (file CWP Quat10) Year 2010.

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Figure 15 Widespread slickensides in the fault gulley are evidence of faulting.

A short introduction to beachrocks, with a report on occurrences on La Palma, Canary Islands

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Introduction

Beachrocks are sediments that have been rapidly lithified by carbonate cement within the intertidal or near supratidal zone to form hard surfaces on what would otherwise be sand or pebble beaches (Vousdoukas *et al.* 2007).

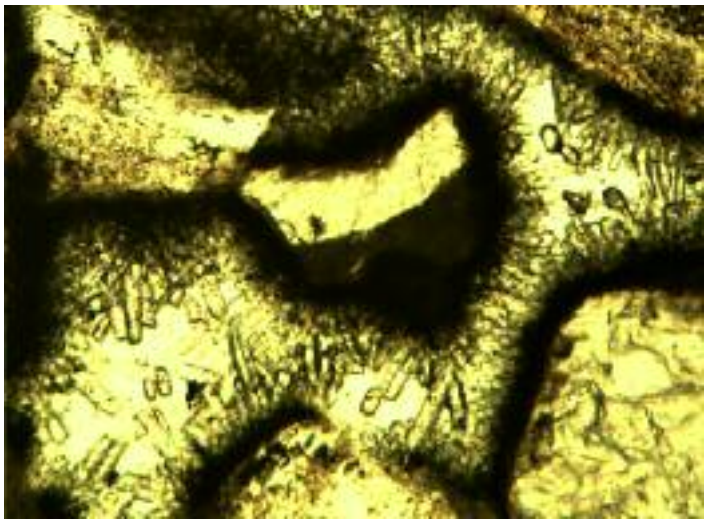
Beachrocks form mainly along tropical or subtropical coastlines, although there are occurrences of beachrock around the Mediterranean and on the coasts of some oceanic islands in the Atlantic. Beachrocks appear to be much less common in temperate regions, although one example in the Outer Hebrides is described in the literature (Kneale and Viles 2000).

Petrography

The grains in beachrocks vary widely, from examples with all carbonate grains to examples where the grains comprise entirely of volcanic material. Examples of beachrock have been described that contain fragments of Roman pottery (Kelletat 2005) and even World War II debris (Frankel 1968). Grains of such recent origin attest to the rapid rate of cementation that may occur in favourable environments.

The nature of the grains appears to have little or no effect on the type or morphology of the carbonate cement. Beachrock cements comprise metastable phases of calcium carbonate, either high magnesium calcite or aragonite (Bathhurst 1975). High Mg calcite (>4% molar Mg) usually occurs in a micritic or stubby, bladed form. Aragonite tends to occur in coarse elongate or acicular (needle-like) crystals that grow orthogonally to the grain boundaries. Controls on cement composition and morphology appear to be poorly understood (Vousdoukas *et al.* 2007). The amount of cement is variable, from very porous beachrocks with just sufficient cement to hold the grains together (Fig. 1) to well cemented beachrocks with little residual porosity (Fig. 2).

Figure 1 Photomicrograph in PPL of a poorly cemented beachrock comprising quartz sand grains cemented by a sparse framework of aragonite crystals — Sharm el Sheikh, Sinai Peninsula, Egypt (field of view c. *2mm).*



Formation of beachrock

There are a number of theories proposed to account for the formation of the beachrock cements, the principal ones are as follows (Vousdoukas *et al.* 2007):

(i) direct precipitation of calcium carbonate from seawater by repeated wetting and drying within the intertidal zone;
(ii) mixing of marine and meteoric waters, which can in some circumstances produce calcium carbonate supersaturation;
(iii) CO_2 degassing, either by biological activity or by increasing pore water temperature as the intertidal zone is exposed to solar warming when emergent.

The impact of beachrocks

Although beachrocks can have value in reducing coastal erosion (Calvat *et al*. 2003), they are generally viewed negatively by both tourists and the local population (Vousdoukas *et al.* 2009). The hard, irregular surface covers what would otherwise be a sand or pebble beach. The beachrock is, in turn, often covered by slippery green biomass (Fig. 3, *overleaf*). Beachrocks thus tends to reduce the aesthetic appeal of a coastal location to all except a few geoscientists. In addition they make access to the sea more difficult and dangerous. Some tourist locations have removed beachrocks or cut access channels through them to allow bathers to get to the sea without mishap.

Beachrocks in the Canary Islands

Occurrences of beachrocks on Fuerteventura have been described by Tietz and Muller (1971). Calvat *et al.* (2003) provides a detailed description of four beachrock locations on the southwest coast of La Palma (Fig. 4, *page 43*), together with a fifth location (Las Zamoras) where the beachrock was destroyed by a

Figure 2 Variably vesicular basalt grains cemented by aragonite to produce a well cemented beachrock with little or no residual porosity — Playa Charco Verde, La Palma (diameter of coin = 20mm).

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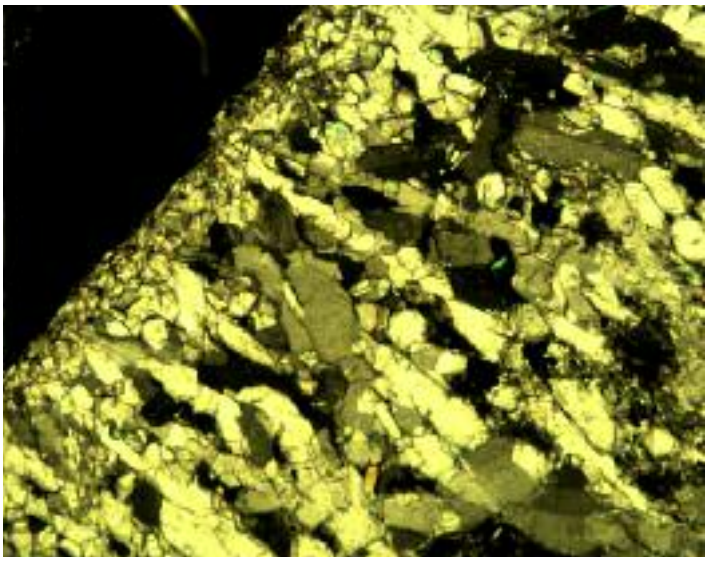
Figure 3 Beachrock exposed at Playa Charco Verde, La Palma. Note the overlying lava flow from the 1585 eruption and the extensive growth of green algae.

storm in 1999. I visited all four remaining beachrock locations during a stay on La Palma in 2010. The beachrock at Puerto Naos appears to have been largely removed for 'beach improvement', while the beachrock at Playa Chica was not well exposed. There were good exposures of beachrock at the remaining two localities; these are described below.

At Echentive the beachrock has developed on top of lavas from the 1677 eruption from the Cumbre Vieja volcano. In places it is partly overlain by a lava flow from the 1971 Teneguia eruption. The beachrock is best exposed around the seaward margin of a saline lake that has developed in the backshore zone. Access is relatively difficult due to growth of slippery biomass.

The beachrock at Playa Charco Verde is well exposed and relatively well preserved, despite the location being a popular beach. The beachrock developed on a platform of Cumbre Vieja lavas with a K-Ar date of *c.* 20ka. In places the beachrock exposures are overlain by lava flows from the 1585 Puerto Naos eruption from the Cumbre Vieja (Fig. 3). The exposures are patchy, in places up to 10m wide, 50m long and dipping seaward at 2–4°. The grains are subrounded to angular, typically

Figure 5 Photomicrograph in XPL of aragonite cement in a beachrock at Playa Charco Verde, La Palma. The dark area in the top RHS is the edge of a basalt grain (field of view c. *2mm).*



5–30mm in diameter and comprising variably vesicular basalt (Fig. 2). The beachrock is well cemented by aragonite. Figure 5 *(opposite)* shows a typical thin section view of the cement, which comprises elongate aragonite crystals that grew orthogonal to grain surfaces.

Acknowledgements

I wish to thank the Lancaster Environment Centre, Lancaster University, for the use of their facilities for the production of Figures 1 and 5.

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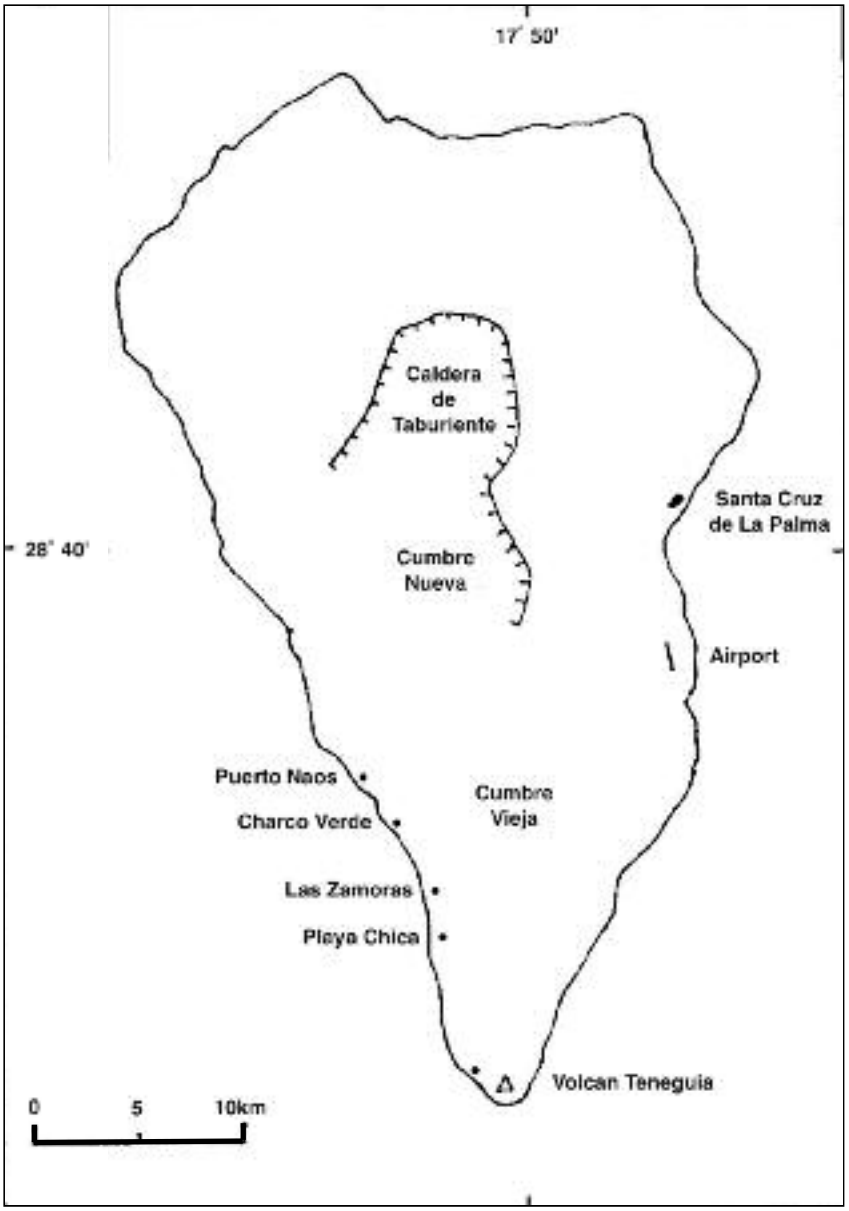

Figure 4 Sketch map of La Palma, showing the location of the beachrock occurrences described by Calvat et al. *(2003).*

[continued from page 32]
'cladogram size, balance and duration', and that comparison of cladograms must therefore be 'made with caution'.

P. D. Wall, L. C. Ivany and B. H. Wilkinson challenge the assumption that marine fossil biodiversity shows a more complete record than do terrestrial fossils environments. Terrestrial successions may in fact show (accurately) an exponential rise in diversity. These authors evaluate the relationships between terrestrial diversity and exposed areas of sediments using, continental outcrops' and generic diversity shown in the Paleobiology Database — and find a high degree of correlation. Correlation is not found to be between habitat area and outcrop size, or between habitat and biodiversity — i.e. there is no common cause for the correlation.

M. J. Benton, A. M. Dunhill, G. T. Lloyd and F. G. Marx contribute a more wide-ranging paper assessing the 'quality of the fossil record', focusing on vertebrates. First they review the assumptions usually made in the use of 'sampling proxies', and show that variations of outcrop area, 'rock availability', rock heterogeneity and fossil content make sampling proxies unable to provide a good estimate of 'what is missing' (the gap between the observed fossil record and reality). They then suggest four approaches to examine the incompleteness of the fossil record: (1) regional-scale studies of geological completeness; (2) regional- or clade-scale studies of sampling completeness (with comprehensive measures of sampling); (3) Phylogenetic and gap-counting methods; and (4) model-based approaches that compare sampling as one of several explanatory variables with measures of environmental change. The authors suggest that palaeontologists 'should simply accept that their data are patchy and incomplete' [i.e. live with it!]

S. E. Peters and N. A. Heim address similar questioning about the reality or bias of observed fossil data. They test the two competing hypotheses: (1) that similarities reflect sampling bias due to 'variable and incomplete sampling' of the fossils; or (2) that similarities reflect 'environmental perturbations' that influence patterns of sedimentation and macroevolution [and that this = a 'common cause']. Using the Paleobiology Database they tested these hypotheses in order to distinguish between geologically controlled sampling bias versus a common cause. They measure 'the intersection of Macrostrat and the Paleobiology Database', and suggest that fossil record patterns are not determined by large-scale geological stratigraphic bias. Rather, linkages between multiple Earth systems drive spatial-temporal patterns in both sedimentation and macroevolution.

J. S. Crampton, M. Foote, R. A. Cooper, A. G. Beu and S. E. Peters report on their study of the fossil record of Cenozoic New Zealand. They state that there is 'increasing evidence' to suggest that bias in the fossil record and actual biodiversity are affected by the same 'drivers'. They explore environmental spatial structuring as 'one component of this common cause hypothesis', and find that spatial structuring of the environment directly influences the quality of the fossil record in Middle Eocene shelf molluscs in New Zealand. But they also found no evidence that environmental spatial structuring was a driver of taxic rates, net diversity or spatial structuring of these molluscs at the scale of their analysis: changes in net diversity are associated with origination rates; and an unknown common cause associated extinction rate and quality of the fossil record. They suggest that tectonic drivers are key in 'the common cause.'

M. Zuschin, M. Harzhauser and O. Mandic try to '[disentangle] palaeodiversity signals' from the biased sedimentary record in the Early to Middle Miocene 'Central Paratethys Sea'. They looked at mollusc diversity and environmental bias across this time boundary. The changes they found over a period of 'less than four million years' developed along a gradient similar to the depth-related environmental gradient. Their study emphasises the need for highly resolved stratigraphic and palaeoenvironmental data.

D. B. Lazarus looks at deep-sea plankton microfossils. He states that there are only a few gaps in the Cenozoic plankton fossil records, and thus that the vast numbers of specimens and core samples provide sufficient evidence to 'integrate macroevolutionary and palaeoenvironmental data'. He reviews documentary problems in the current records (incomplete taxonomies, incomplete coverage of the total diversity and inaccurate 'age models'), but concludes that additional work on and refinement of these issues will make the record adequate for the task.

G. T. Lloyd, A. B. Andrew and J. R. Young study coccolithophores to address these issue of fossil record bias and deep-sea rock quantities. Their core samples from the North Atlantic, Mediterranean and Caribbean show 'an exponential rise [in species occurrence] towards the Recent'. Over this same time period species-richness grew linearly while genus-richness showed an initial sharp increase, then a slow decline. They conclude that the coccolithiphore deep-sea record has a rock record bias.

A. D. Baronsky, M. A. Carrasco and R. W. Graham contribute a paper comparing late Quaternary 'collateral mammal diversity loss' and megafaunal extinctions — *c.* 30Ma to 500 years ago in 10 biogeographical provinces of the USA. They conclude that both mammalian groups remained stable until both 'crashed near the Pleistocene–Holocene transition'. Megafauna extinction amounted to *c.* 21% of pre-crash species, with collateral biodiversity loss (large mammal diversity fell an additional 6–31% above extinction loss; small mammal diversity fell 16–51%). However, few small mammals actual went extinct. They suggest that these results imply that collateral loss effectively doubled regional diversity loss, and that these results for the recent past have implications for the prognosis for mammal diversity in the USA and worldwide.

R. B. J. Benson and R. J. Butler uncover diversification of marine tetrapods and explore ecological influences on geological sampling bias. Their statistics indicate that shallow marine vertebrate biodiversity [data] are driven by changes in the extent of continental flooding, and only weakly by uneven sampling. They say that this conclusion is 'congruent' with the evidence for shallow marine invertebrates, implying that there was a 'common cause' for shallow marine biodiversity fluctuation. In contrast, their evidence for open ocean tetrapod diversity is distorted by 'temporal heterogeneity in fossil record sampling' — i.e. is unrelated to continental flooding. In other words ocean-dwelling plesiosaur, ichthyosaur and sea turtle species avoided periodic extinctions driven by major marine regressions in the Late Triassic and across the Jurassic–Cretaceous boundary. The evidence shows that open-ocean taxa declined before the end-Cretaceous extinction, while shallow-marine taxa continued to diversify at this time owing to sea level increase.

P. Upchurch, P. D. Mannon, R. B. J. Benson, R. J. Butler and M. T. Carrano use 'the Dinosauria' as a case study of geological and anthropogenic sampling control of the fossil record. They studied

[continued on page 50]

A geological traverse along the Sawdde Gorge and over Black Mountain, Carmarthenshire

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Introduction

Black Mountain is the name of a range of hills rising to a height of over 600m in eastern Carmarthenshire that lies within the Fforest Fawr Geopark in the western part of the Brecon Beacons National Park. The name Black Mountain should not be confused with The Black Mountains to the north of Abergavenny. One of the most celebrated travellers to cross the Black Mountain was George Borrow in 1854. He describes in his book *Wild Wales* (Barrow 1955) how he stayed overnight in Llangadog and the following day walked over the mountain to Gutter Vawr (present day Brynamman). "Leaving Llangadog I pushed forward. The day was now tolerably fine. In two or three hours I came to a glen, the sides of which were beautifully wooded. On my left was a river, which came roaring down from a range of lofty mountains right before me to the south-east. The river, as I was told by a lad, was the Sawdde or Southey, the lofty range the Black Mountain".

The traverse described in this article follows the A4069 road for *c.* 20km along the Sawdde Gorge and over Black Mountain from to Llangadog in the north to Brynamman in the south (Fig. 1) along much of the route followed by Borrow. All the beds strike approximately east–west across the region and they dip south so that the oldest rocks are in the north. They range in age from Lower Silurian (Llandovery Series) to Upper Carboniferous (Pennant Sandstone Formation) (Fig. 2, *overleaf*). The Afon Sawdde rises on the northern slopes of Black Mountain in the cwms (large glacial eroded hollows) of Bennau Sir Gaer. It cuts a spectacular gorge as it flows north-westward to its confluence with the Afon Tywi near Llangadog. This itinerary follows up through the stratigraphical order of the rocks as we move southward along the A4609 through the gorge and onward over Black Mountain.

The classic Sawdde river section (Fig. 3, *page 47*) exposes a continuous succession of Silurian strata southward from Bont Fawr [SN715260], where mudstones of the Upper Llandovery Series rest unconformably on Llandeilo Flags (Middle Ordovician). Upstream the river cuts through a sequence of sandstones and mudstones, including the Sawdde Siltstone, the Sawdde Sandstone and the Ffinnant Sandstone formations of the Wenlock Series. The disused quarries of Cwar Glâs (Locality 1) adjacent to the river provide one of the best inland exposures of the succeeding formations of Ludlow age in the region. Some 300m south of the quarries at Pont-au-llechau, the Tilestones Formation outcrops at the base of the Old Red Sandstone. The succeeding Raglan Mudstone Formation (Přídolí Series) is exposed in the bed of the Afon Sawdde for *c.* 1km immediately above Pont-au-llechau. It is notable that the Pickard Bay and Townsend Tuff form stratigraphically important marker beds near the top of this formation outcropping *c.* 200m north of Turkey Cottage [SN733237]. These distinctive ash fall tuffs can be traced across into Pembrokeshire, where they occur in numerous coastal localities (see 'Field observations in the Old Red Sandstone of the South Pembrokeshire coastline', *OUGSJ* **29** (1) 2008, 10–15).

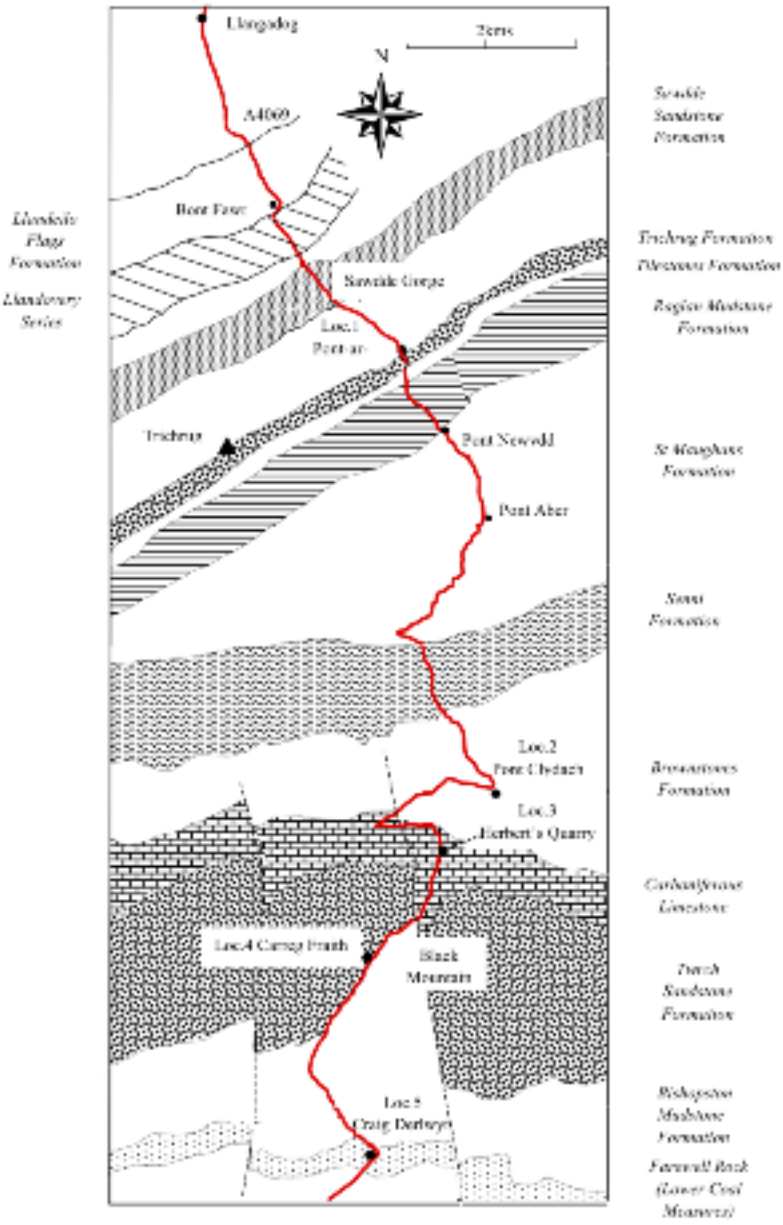

Figure 1 Map of Black Mountian.

The rocks of the Sawdde Gorge are highly significant because they provide a continuous record from which to interpret the changing palaeoenvironment in south-west Wales during the Silurian Period. By late Llandovery times a broad shelf sea extended along the margins of the Welsh Basin where shallow marine sediments were being laid down. The Wenlock beds coarsen upward containing cross-stratified sandstones that show evidence of wave and current action (Ffinnant Sandstone Formation). By Ludlow times deeper water offshore conditions prevailed (Tresglen Formation), followed by marine regression as a northward prograding delta deposited sands and conglom-erates (Trichrûg Formation). The transition from

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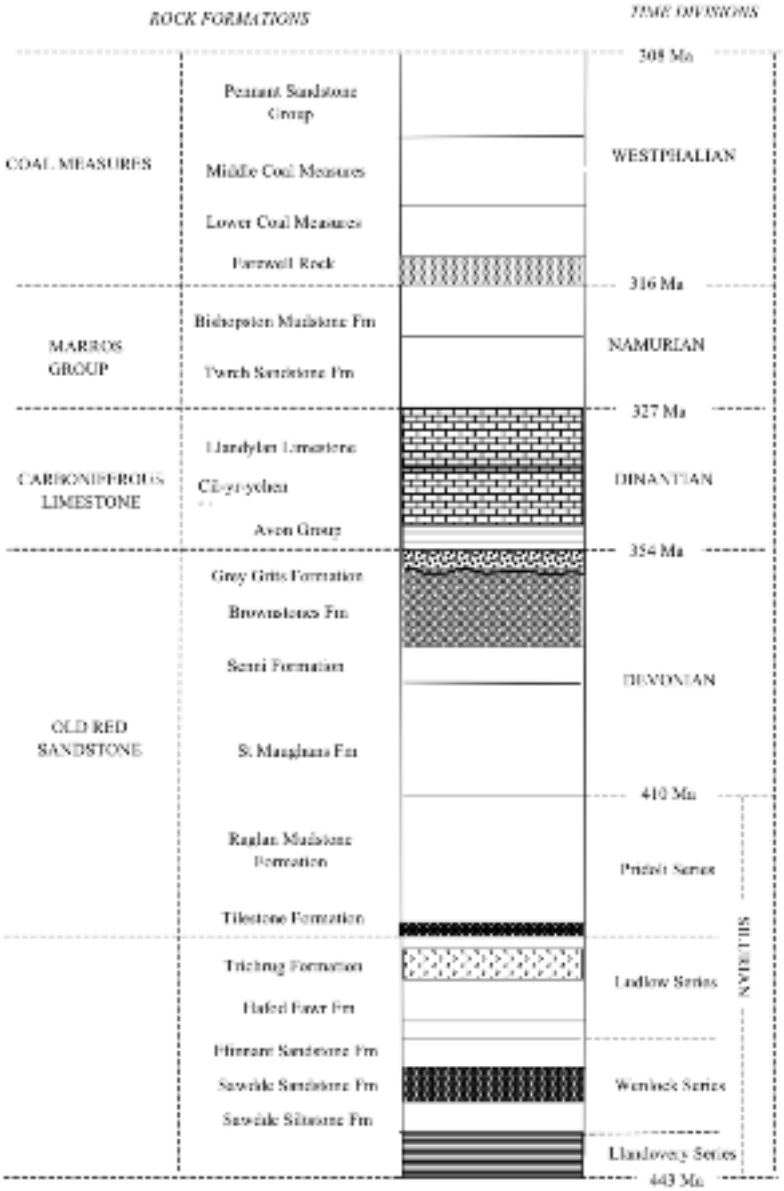

Figure 2 Stratigraphic column for Black Mountain.

marine to continental conditions was taking place and by Přídolí times red mudstones with calcrete profiles were being deposited by meandering rivers flowing across wide alluvial plains (Raglan Mudstone Formation).

The Sawdde gorge section ends at Pont Newydd [SN737235] from where the A4069 climbs up the slopes of Black Mountain with its undulating stepped topography reflecting the outcrops of hard bands of sandstones and limestones interbedded with softer rocks such as shales and marlstones. The Carboniferous Limestone forms a distinctive scarp feature on the northern edge of Black Mountain and the karstic topography of the limestone includes caves and swallow holes although much of the original surface has been removed by quarrying. In the 19th century limestone was in demand for both agricultural fertilizer produced in the quarry lime kilns and for iron production on the South Wales coalfield. The Amman Valley ironworks in Brynamman and the Ynyscedwym works at Ystradgynlais in the Upper Tawe Valley used local limestone as a flux in the blast furnaces which were fuelled by local anthracite. When iron making was superseded by steel manufacture, the Twrch Sandstone was quarried for silica to make refractory bricks in the steel furnaces.

The traverse covers the region lying to the south of the Twyi anticline, a major Caledonian fold structure trending north-east–south-west along the Tywi Valley. The Sawdde Gorge cuts through the south-east flank of the anticline, where the steeply dipping Silurian strata are exposed along the sides of river. However, the relatively simple geological structure is complicated by the Carreg Cennen Disturbance, a narrow zone of thrust faulting and intense folding that runs west-south-west–east-north-east on the northern margin of the Black Mountain. Along the line of this tectonic disturbance there are several faulted outliers of Carboniferous Limestone surrounded by Old Red Sandstone, including the Carreg Cennen Castle outlier described in Locality 6. The Carboniferous rocks of Black Mountain dip southward to form the northern edge of the South Wales coalfield. This broad synclinal structure, with its axis orientated east–west, is the product of the northward propagation of the Variscan Front during late Westphalian times.

The following localities provide an insight into the geology of this scenically attractive part of eastern Carmarthenshire.

Locality 1 Pont-au-llechau [SN728245]

The narrow gorge of the Afon Sawdde is accessible with care for most of its length. There is a rough track from Bont Fawr following the east side of the gorge. However, there are few suitable parking places except at the picnic site on the A4069 at SN717255 and at the hamlet of Pont-au-llechau. The latter is recommended for a visit to the old quarries of Cwar Glâs [SN727247]. Cross the bridge in Pont-au-llechau and follow the narrow road alongside the river for *c.* 300m to the quarries (Fig. 4, *opposite*), which are cut into the Hafod Fawr Formation, formerly known as the Black Cock Formation (Ludlow Series, Upper Silurian). The beds dip at *c.* 70° south-east and include mudstones, sandstones, gritstones and conglomerates; there are also some fossiliferous calcareous beds and rottenstones (‘gingerbread’ bands) that yield small bivalves, gastropods and brachiopods. The thick purple sandstones in the east face of the southern quarry are referred to as the Cwar Glas Member. There are some excellent well-preserved ripple marks on the steeply inclined bedding planes (Fig. 5, *opposite*), which together with the presence of coarse-grained, cross-stratified sandstones and conglomerates, provide evidence of the shallow water conditions under which these sediments were deposited. The ripples were formed when the sediment was in a soft, wet condition, possibly in a tidal estuary or in a shallow marine environment.

After leaving the quarry descend to river level near the confluence with the Afon Meilwich, a small tributary entering the Sawdde from the south-west. This is *c.* 200m north of the bridge at Pont-au-llechau, where the red pebbly sandstones of the Trichrûg Formation are exposed. These beds mark the beginnings of the change to continental conditions that heralded the beginning of the deposition of the Old Red Sandstone. These relatively hard resistant beds form the ridge of high land extending 3km to the west known as Trichrûg [SN699230]. Continue along the river towards the bridge where the Tilestones Formation outcrops. The Tilestones are the lowest beds of the Old Red Sandstone and were formerly used for

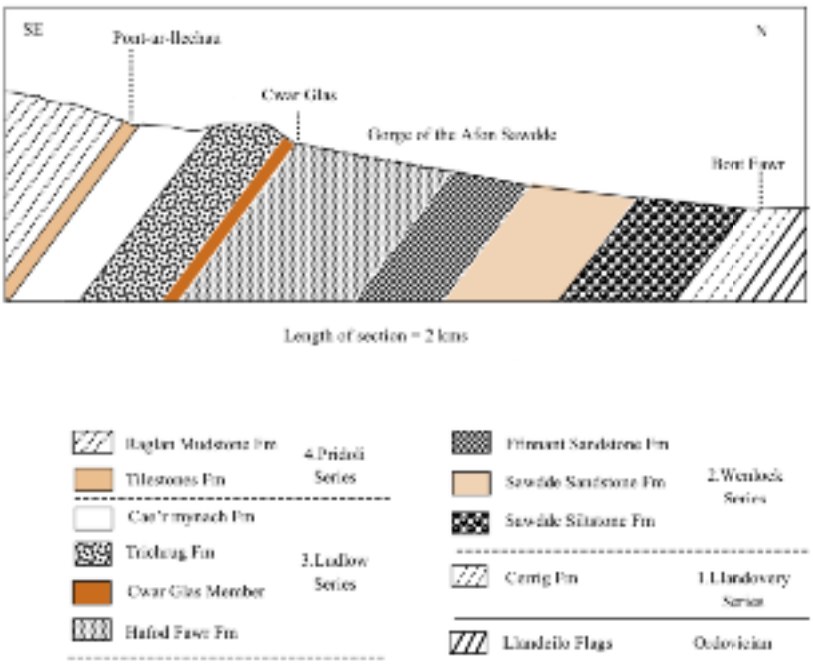

Figure 3 Sawdde Gorge Section.

roofing local houses and barns, as they are flaggy, micaceous sandstones. Significantly the Welsh name of the village means 'bridge on the slates'.

Locality 2 Pont Clydach [SN739197]

From Pont-au-Llechau drive for *c.* 6km along the A4069, crossing the red sandstones and mudstones of the Lower Old Red Sandstone and climbing gradually up to the sharp bend in the road at Pont Clydach. There is a small lay-by here and it is worth walking up the side of the small valley to examine the conglomerate boulders in the bed of the stream. These boulders are derived from the Grey Grit Formation (Upper Old Red Sandstone) higher up the hillside. At this locality several small waterfalls occur where the stream crosses harder bands of sandstone. There is a good exposure of the Brownstones on the roadside *c.* 300m west of Pont Clydach, but it is difficult to park near by. The Brownstones are red micaceous sandstones with some shales and marls that form much of the northward facing scarp face of the Black Mountain.

Locality 3 Herbert's Quarry [SN733187]

From Pont Clydach continue up the road around the hairpin bend to the car park and view point at SN731193 (Fig. 6, *overleaf*). Just north of the car park you will see a prominent bench-like feature dipping south at a low angle. This is the Grey Grit, a coarse quartz conglomerate containing pebbles of vein quartz and red jasper that forms the topmost unit of the Old Red Sandstone (Fig. 7, *overleaf*). The outcrop of the Grey Grit crosses the road but is seen to die out completely *c.* 100m west, where it is cut by the Cwmllynfell Fault running north–south through a grassy gully. Beyond the fault line to the west are several old quarries [SN726192] in the Llandylan Limestone, which is overlain by the Honeycomb Sandstone. This is a buff-coloured, poorly consolidated calcareous rock honeycombed with sand filled hollows.

Clearly there appears to have been a lateral displacement of the strata by some 400m, due to the downthrow of the Cwmllynfell fault to the west. The fault has displaced the Grey Grit outcrop and we need to look to the north of the hairpin bend to see its westward continuation. The conglomerate outcrop can be seen to run sub parallel to the minor road (signposted to Trapp) that joins the A 4069 just north of the hairpin bend. The Avon Group shales form the low ground at the foot of the overlying limestone and a series of springs emerge at the junction of the shale and limestone producing several calcareous tufa waterfalls.

Next proceed along the main road past the memorial stone for *c.* 500m to the picnic site and car park at SN733188. Immediately north of this car park is a lane which leads to several disused quarries in the Dowlais Limestone Formation

Figure 4 Sketch of Cwar Glas.

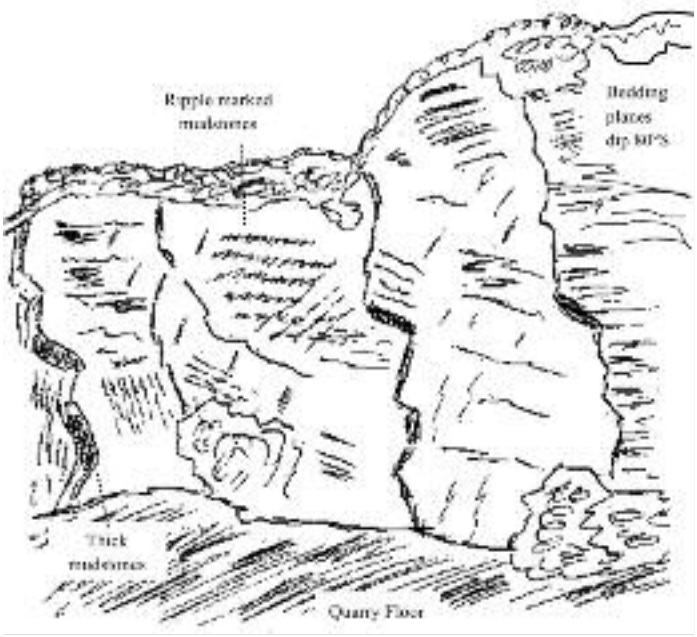

Figure 5 Ripple marks on the steeply inclined bedding planes, Cwar Glas North.

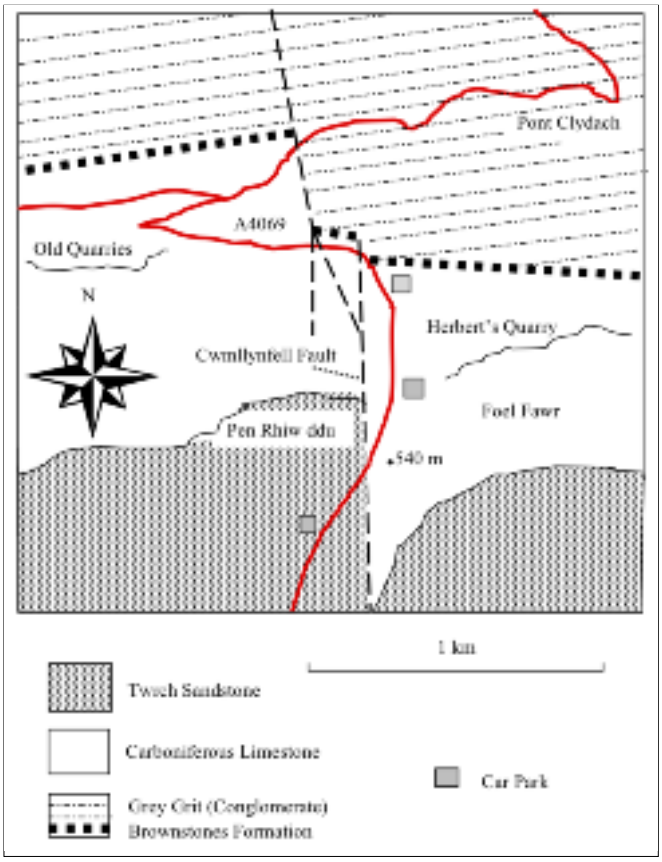

Figure 6 Map of Black Mountaint summit.

(Fig. 8), locally known as the Llandyfan Limestone (Holkerian Stage S2 subzone). These quarries were used to provide limestone for flux in iron smelting on the South Wales coalfield and also the limestone was burned in the quarry kilns to produce lime for agricultural fertilizer. Needless to say, the stone was also used for the construction of local stone walls, cottages and farm buildings. Brachiopods such as the zone fossil *Composita [Seminula] ficoidea* can be found particularly in some of the partially burned blocks of limestone near the old limekilns. Herbert's Quarry is a large shallow quarry *c.* 300m from the car park that has some fine bedding planes with rectilinear joint patterns. The limestone dips gently southward and this is

Figure 8 Disused quarry in the Dowlais Limestone Formation, locally called Llandyfan Limestone.

Figure 7 Grey Grit, the topmost unit of the Old Red Sandstone.

admirably demonstrated by water accumulating along the quarry wall, acting as a spirit level (Fig. 9). About 100m north of the quarries there is a small scarp that overlooks a strip of low marshy ground. This marks the outcrop of the Avon Group (formerly the Lower Limestone Shales) at the base of the Carboniferous Limestone.

Locality 4 Carreg Fraith [SN720172]

Continue up the A4069 to where the road reaches its maximum elevation on Black Mountain, where there is a car park at SN 730183. Just before the car park you will cross the line of the Cwmllynfell Fault, which runs in roughly a north–south direction (Fig. 6). Here the Carboniferous Limestone is exposed on the east side of the fault in juxtaposition to the Twrch Sandstone on the west side. Several swallow holes can be seen aligned along the fault zone. From the car park walk southward to the small roadside quarries in the Twrch Sandstone Formation (formerly the Basal Grit), which is exposed on the eastern flank of Garreg Fraith. The light grey rocks are almost pure quartzite with a silica content of *c.* 98%, which made them suitable for the manufacture of refractory bricks for lining the furnaces in steel works. These hard resistant beds of Twrch Sandstone form the crest of Black

Figure 9 Sketch of Herbert's Quarry.

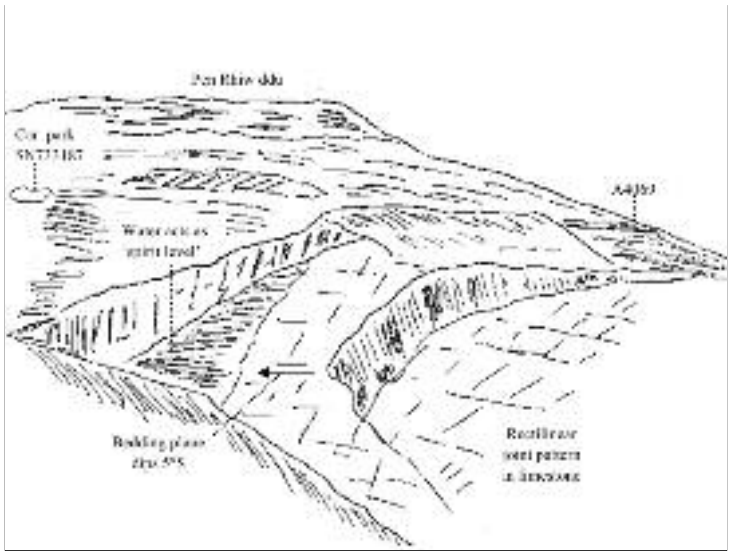

Figure 10 Craig Derlwyn, where the Gastrioceras subcrenatum *marine band marks the base of the Lower Coal Measures occurs immediately above the shales.*

Mountain, with bare gritstone tors and boulder-strewn slopes dominating the summit area. The tors probably formed under periglacial conditions during late Devensian times. Much of the surrounding moorland lies on thick peat deposits and where exposed, the peat gives a dark aspect to the hills, hence the name Black Mountain.

Locality 5 Craig Derlwyn [SN725158]

This locality lies *c.* 2km north of Brynamman on a sharp bend in the road at Craig Derlwyn. Parking is available alongside the old quarries on the left hand (west) side of the road. The quarries are cut into the basal member of the Lower Coal Measures known as the Farewell Rock, so called because it lies below the lowest workable coal seams. It is worth noting that the outcrop of the Farewell Rock is diachronous, as to the west in the Pembrokeshire coalfield it occurs at the top of the Marros Group, formerly known as the Millstone Grit. At Craig Derlwyn the quarries expose massively bedded siliceous sandstones with interbedded flagstones and shales. The sandstones are strongly jointed and display well developed spheroidal weathering. They were laid down under fluvial conditions as meandering rivers flowed southward off the Wales-Brabant landmass depositing sediment in large deltas fringing a southern sea.

The Farewell Rock is underlain by the Bishopston Mudstone Formation (formerly the Middle Shales) belonging to the Marros Group. The junction between the two appears *c.* 300m north-west of Craig Derlwyn at SN720161 (Fig. 10). The *Gastrioceras subcrenatum* marine band that marks the base of the Lower Coal Measures occurs immediately above the shales. Bands of quartzite become increasingly common near the base of the shales.

From the vantage point of Craig Derlwyn you can look south across the low ground beyond Brynamman that is formed by the Lower and Middle Coal Measures on the northern margin of the South Wales coalfield. Former opencast workings occupy much of the area immediately south of Brynamman and beyond lies the scarped edge of the overlying Pennant Sandstone, which forms a much dissected plateau feature.

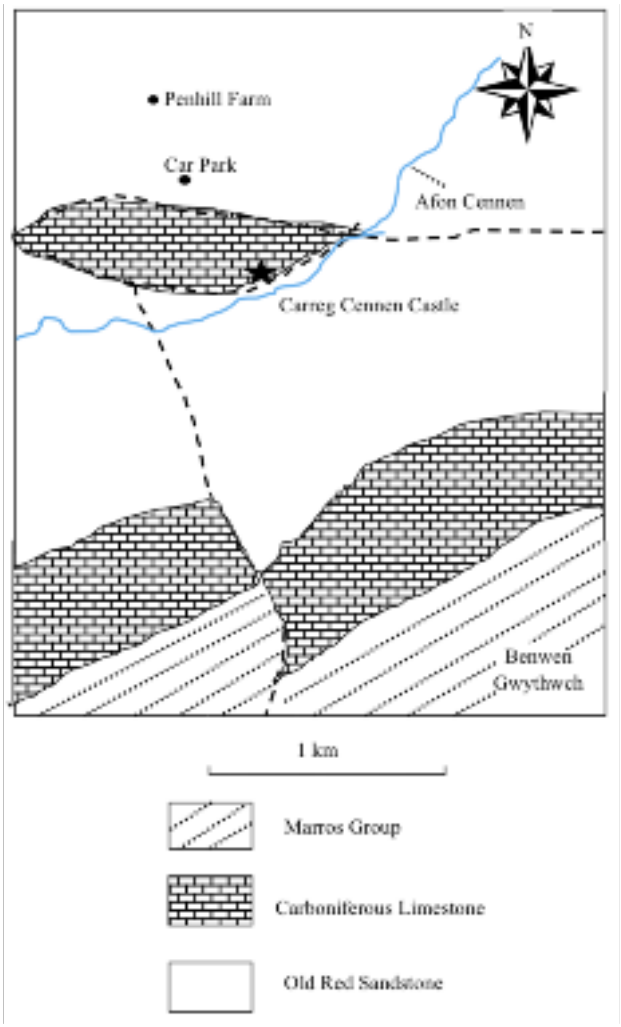

Figure 11 Map of Carreg Cennen.

Locality 6 Carreg Cennen Castle [SN668191]

This locality lies *c.* 8km west of the summit of Black Mountain and it can be reached by travelling along the narrow winding minor road leading from the hairpin bend [SN725193] on the A4069. The road is signposted to Trapp, which is the nearest village to the castle. There is a car park, toilet facilities and a shop on the north side of the castle.

The castle stands on a prominent hill on the north side of the Afon Cennen valley. It was the power base of the princes of Deheubarth in the 12th and 13th centuries, but it was later captured by the English and was subsequently abandoned. Structurally the castle hill is formed by a thrust slice of Carboniferous Limestone (Cil-yr-ychen Limestone), sandwiched between the St Maughans Formation to the north and the Senni beds to the south. The limestone is a faulted outlier (Fig. 11) of the main outcrop of Carboniferous Limestone that forms the northern edge of Black Mountain. As you walk from the car park to the castle the steeply bedded red marls are exposed in the track, and just beyond the entrance gate is an elongated depression that marks the line of the basal thrust. The limestone cliff on which the castle stands is much shattered due to the effects of the northward-directed thrust movement (Fig. 12, *overleaf*).

This Carreg Cennon Disturbance is a major thrust zone that can be traced south-westward across Carmarthen Bay to the Johnston Thrust in Pembrokeshire. The Afon Cennen flows through a deep wooded gorge below the castle cliffs (Fig. 13, *overleaf*) following the line of the thrust fault zone.

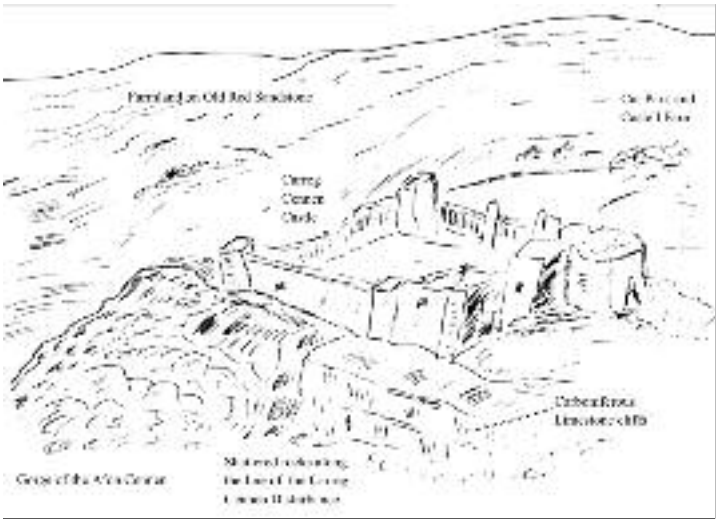

Figure 12 Sketch of Carreg Cennen Castle.

Figure 13 Afon Cennen and deep, wooded gorge below Castle Carreg Cennen cliffs.

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Maps

British Geological Survey 1:50,000 Sheet 230 Ammanford

British Geological Survey 1:50,000 Sheet 212 Llandovery

Ordnance Survey 1:50,000 Sheet 160 Brecon Beacons

Ordnance Survey 1:50,000 Sheet 159 Swansea and Gower (includes the Carreg Cennen area)

Outdoor Leisure Map 1:25,000 Sheet12 Brecon Beacons — west and central areas

Author

John Downes is the author of *Folds, Faults and Fossils: Exploring Geology in Pembrokeshire* (2011; Llygad Gwalch, Pwllheli)

[continued from page 44]

847 dinosaur species. Using several 'sampling metrics' they conclude that 'different metrics often capture the same signal'. They show that regional sampling correlates well with regional diversity, but more weakly with global diversity. Sauropodomorphs diversified in the Late Triassic; in contrast, theropods and ornithischians did not diversify until the Early Jurassic. Diversity in all groups continued in the Jurassic, but many groups became extinct across the Jurassic–Cretaceous boundary. There was some species diversity recovery during the Cretaceous, but only sauropods show a long-term upward trend. The Campanian–Maastrichtian sauropod 'diversity peak' appears to be 'largely a sampling artefact'.

As said, these papers are technical. But there is a richness of data in this anthology from which each reader can explore his or her particular palaeontological interest. Each article includes numerous diagrams and charts to summarise the statistics; and has its own extensive list of referenced sources. The book concludes with a seven-page index.

— David M. Jones, BA UC Berkeley, MA, PhD UC London, BSc Open, OUGS Journal Editor

OUGS Symposium 2011 field trip to Callander, Trossachs, central Scotland

(Led by Dr Mike Browne, BGS, Edinburgh, retired)

Tom Miller, OUGS Walton Hall

Introduction

This field trip took place on Saturday, 20 August as part of the OUGS Stirling Symposium 2011. It was led by Dr Mike Browne (retired), formerly of the BGS, Edinburgh (Fig. 1). Callander is located just south-east of the Highland Boundary Fault (HBF), which crosses Scotland from Aran to Stonehaven. We were to visit several locations, taking in local landmarks, glacial features and exposures of the rocks close to the HBF on the south side.

Figure 1 Dr Mike Browne, Callander Moraine, Drumdhu Wood (© Don Cameron 2011).

Location 1 Callander esker

We got off the bus in Callander at 'the Roman Camp' on the main A84 A narrow path led onto a wooded ridge, *c.* 4–5m high. This, John explained, was an esker that marked the winding path of a glacial river that had passed beneath the glacier and reached the bedrock beneath. It had formed *c.* 11,600 years ago at the end of the last ice age (Loch Lomond Stadial). It had wound its way down-slope toward the snout of the glacier, leaving a distinctive sequence of river lag and sediments, now standing proud of the land surface as a long, sinuous ridge.

In earlier times, the local people had assumed the esker to be an earthwork, associated with a Roman camp thought to be near by. The esker deposit, visible amid the roots of the trees, seemed to consist mostly of large, rounded cobbles, gravel and sand, suggesting a periodically powerful outwash flow.

John led the group along the esker and then down to the bank of the River Garbe Uisge, south and east of Callander. We followed the north bank of the river, towards its confluence with the Keltie Water, then turned north-east up Geisher Road and back to the A84.

Location 2 Callander moraine

This was a wooded area of high ground north of the A84, called Drumdhu Wood. This high ground was a section of the terminal moraine formed by the glacier. A recent attempt by property developers to obtain planning permission to build housing to the west of the moraine has been resisted, with active participation by Dr Browne as a geoconservationist.

We followed the crest of the moraine northward. In profile, it was significantly steeper on its west side, which John explained was the steep west flank in contact with the ice slope.

Coming off the moraine ridge, the path led beside the workings of the Auchenlaich Quarry. We noted sundry large boulders in the restored quarry land to the east. Passing north of the golf course, we could see a high, south-facing escarpment that had formed in the rocks not far south of the HBF line (Fig. 2).

The clearly visible, steeply dipping, almost vertical, strata of the escarpment are the Lower Devonian rocks that face directly onto the HBF, out of sight to the north.

Figure 2 South-facing Callander Craig, an outcrop of one of the now-vertical resistant conglomerates (© Don Cameron 2011).

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Location 3 Former conglomerate quarry, Murchison Road

We returned to town along Livingston and Murchison avenues. Along Murchison Avenue there are some old quarry workings in the roadside bank, revealing quartz-bearing, Lower Devonian A'Chriannach Conglomerate. This had been quarried and used extensively in the building of the town. It was thickly bedded and steeply dipping (Fig. 3). Other examples can also be seen from afar, in the high escarpment to the north.

A right turn off Murchison Avenue into Bracklinn Road leads uphill through woodland to Bracklinn Falls, our next location.

Location 4 Bracklinn Falls

These falls are a much visited natural tourist attraction.

All the strata at the falls are part of the Lower Devonian Bracklinn Falls Conglomerate Member. This consists mostly of beds of conglomeratic sandstone, sandstone and mudstone beds. The beds are sub-vertical, and are younger on the downstream side (Fig. 4). They do not appear to be desert-fluvial, as they show no evidence of channels; rather, they are laterally unconfined, as though they were overbank deposits, or were laid down in a wide estuary or lake. Here and there, rippled surfaces can be found, and there are also worm burrows in the finer beds (Fig. 5, *opposite*). Here we remained on the Laurentian side of the HBF, and so did not see anything of the Dalradian geology.

Figure 3 (left) Quartz-bearing, Lower Devonian A'Chriannach Conglomerate — OUGS Webmaster Stuart Swales for scale (© Don Cameron 2011).

Figure 4 (below) Lower Devonian Bracklinn Falls Conglomerate (© Don Cameron 2011).

Figure 5 Worm burrows in the finer beds — author Tom Miller for scale; detail below (© Don Cameron 2011).

Book review

Park, Graham 2012 *Introducing Tectonics, Rock Structures and Mountain Belts*. Edinburgh: Dunedin Press (ISBN 978-1-90671-627-4; paperback, xii + 132pp; £14.99)

Graham Park, Professor Emeritus of Tectonic Geology at the University of Keele, has written two books for this Dunedin Press series introducing the main fields of the earth sciences. His *Introducing Geology — A Guide to the World of Rocks* (2nd edn) was published in 2010.

The series also includes *Introducing Palaeontology — A Guide to Ancient Life* (2010), *Introducing Volcanology — A Guide to Hot Rocks* (2011), I*ntroducing Geomorphology — A Guide to Landforms and Processes* (2012) [reviewed by Bob Morley in *OUGSJ* **33** (1) 2012 — Ed.], *Introducing Oceanography* (2012) and *Introducing Meteorology — A Guide to the Weather* (2013). Together these books form a useful survey of the earth sciences in sufficient detail to serve as reviews for second-level geology students.

Introducing Tectonics … does just what its cover promises. The Preface begins by stating that the book is not meant to be a textbook — and indeed it isn't. Rather, Park introduces the reader to the major concepts of rock structures, tectonic movements of rocks and their mechanics, and the processes of mountain building or orogenics. The book avoids complex physics and mathematics, the author deciding instead to use geometry at the level necessary to understand the concepts of tectonics instead of involving (and possibly discouraging him or her) the reader in equations and physics.

A short Introduction explains the key terms to set the scene for the rest of the book. Park begins with 'tectonic' and 'structure', 'kinematic' and 'dynamic' models of rock deformation, and then gives brief introductions to each of the chapters that follow, and their principle terms. Every 'key' term is set in blue, and every term or phrase so set is included in the Glossary at the end of the book. This use of blue for key terms / glossary terms prevails throughout the book. In the Glossary, each term or phrase is defined for easy reference; terms used within a glossary definition that are themselves glossary terms, are set in **bold** in the glossary definition, alerting the reader that the emboldened term is itself further defined in the Glossary.

As a long-time book editor I normally find such 'frills' annoying and a little distracting. But here I admit that the use of them makes sense in terms of helping the reader navigate quickly between pages when necessary. Nothing is more annoying that 'knowing' that you have read the definition of a term earlier in the book but now must spend time finding it again, or in looking it up in the index and then going to each index page entry until you find the primary definition to refresh your memory. The differentiation of text word colours is also essential in this book, as there *is* no index, for the Glossary serves that purpose, although bereft of page numbers.

The chapters that follow are: 2 Large-scale Earth structure; 3 Plate tectonics; 4 Deformation, stress and strain; 5 Fractures, faults and earthquakes; 6 Folds and folding; 7 Fabric; 8 Igneous intrusions; 9 Structural effects of gravity; 10 Tectonic interpretation of orogenic belts; 11 Examples of modern orogenic belts; and 12 Ancient orogenic belts. Thus the chapters follow a logical sequence, proceeding from structural concepts and dynamics, through interpreting the evidence of rock structures, to historical geology. The sequence of chapters 2 to 9 is also logical, although I wonder if chapter 7 Fabric should not have come earlier, to explain 'rock structure' before the chapters on major rock movements and ways of deforming the fabric. However, it cannot be easy to decide which concepts to tackle first in such a book, and indeed it is not absolutely necessary to read the chapters in their numerical sequence, unless you are new to the subject rather than a student who has done the OU second-level geology course or equivalent — in which case perhaps Park's *Introducing Geology* might be a recommended book to read before *Introducing Tectonics*. Also, Park feels that the reader needs to understand the model of plate tectonics and the evidence for it before looking at tectonics on smaller scales, and in this I think he is right — thus chapters 2 and 3.

There is also an Appendix preceding the Glossary. This brief end-matter (3 pages) comprises 'tables' with extended captions giving the reader basic, useful geological information that helps put the rest of the book into Earth history context: Geological Time, and a table each for basic classifications of igneous, sedimentary and metamorphic rocks.

The book itself is handsomely produced, in a larger format than, for example *Introducing Geomorphology* or *Introducing Palaeontology* — it measures 220mm × 220mm, with a sturdy card cover wrapped around sewn and gathered page sections, and with front and back in-folding flaps. Without an index, the designers have adopted a somewhat peculiar method of pagination: instead of a page number on each page, the upper right corner of each right-hand page has a mountain vignette with the chapter number in white and, below it, the page numbers for the facing left- and right-hand pages of the spread, divided by red line framing. Instead of conventional running heads of chapter names, each right-hand page also has a 'running head' set at a right angle running 'up' the page margin. These design elements take a little getting used to in my opinion. While not detrimental to the book's content, it remains true that the design of a book can help or hinder the reader to navigate, and I do find *these* unconventional design frills unhelpful.

Much more importantly, however, is the writing. Park writes clearly and convincingly, moving from simpler to more complex topics in mostly short sentences using ordinary words (except, of course, for the necessary geological vocabulary, which is the point of the book!). This is refreshing.

Equally important, the design and use of numerous diagrams, cartoons, charts, annotated photographs, and tinted tables and panels is excellent and complementary to the text. Without exception the illustrations (figures) are clear, well drawn, well reproduced and nice to look at. The figures are numbered separately for each chapter and there are 107 of them.

The book ends with a 'Further reading' list. This includes five books, all of which are textbooks. The reader who wants more will surely find it in these. Also listed are Robert Butler's *Virtual Seismic Atlas* (Univ. Leeds Earth Science Dept) and the same author's useful web-based resource *Learn Structure*.

Finally, Park invites the reader to explore: "put structural geology into a search engine and see what comes up!" — surely a good recommendartion.

I therefore recommend this book to OUGS Members who want a good revision and to students who want a useful, extra non-text 'text' to use alongside their undergraduate text, and with which to revise.

— David M. Jones, BA UC Berkeley, MA, PhD UC London, BSc Open, OUGS Journal Editor

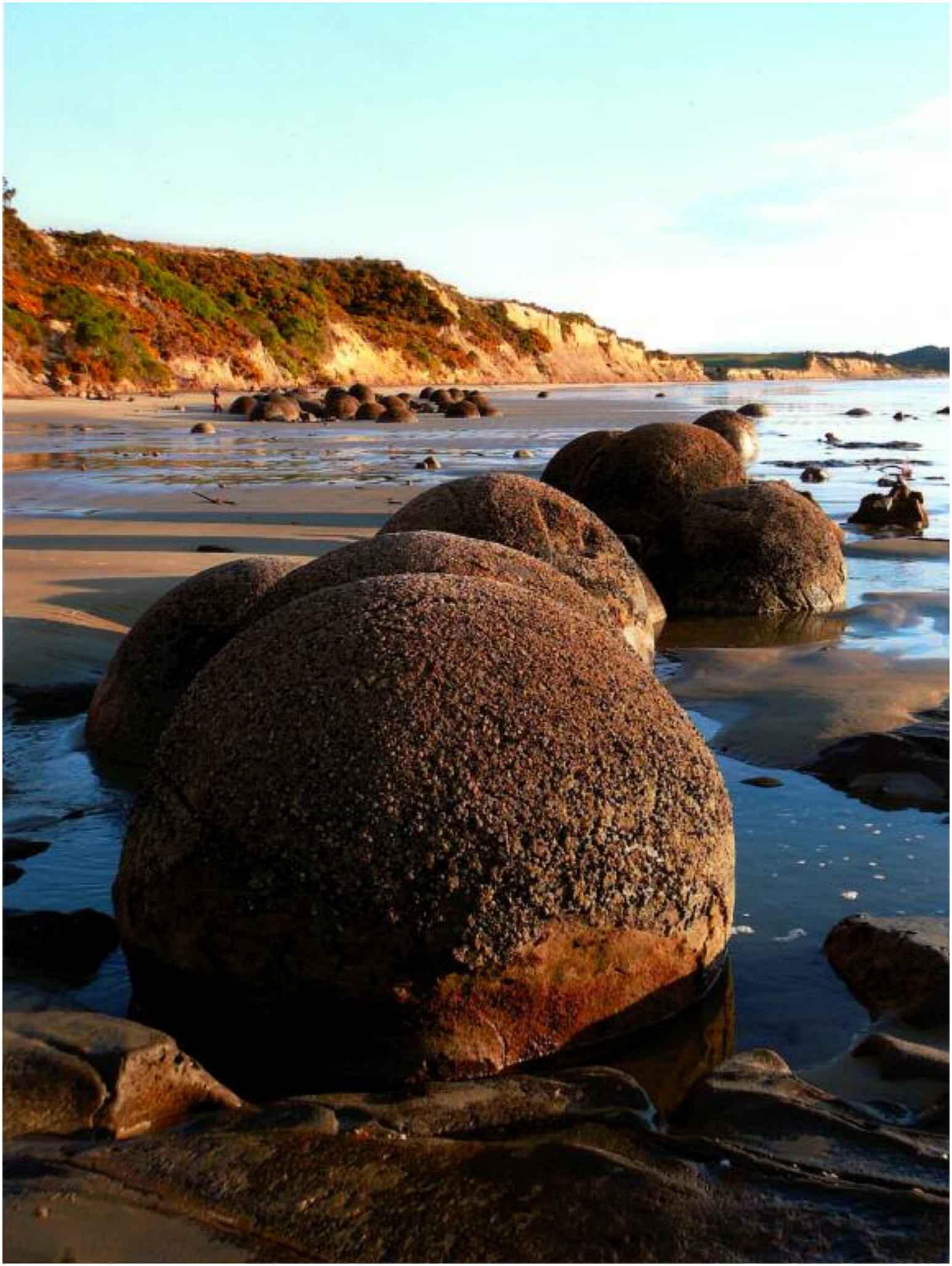
Moyra Eldridge Photographic Competition

Winners for 2012

Geological Structure or Feature and Popular Vote: Alison Chetwin, East Midlands Branch
Bude Formation *(overleaf, left)*
This photograph was taken from the beach, just to the north of Bude in Cornwall. It is part of the Bude Formation, showing sandstone and mudstone intervals deposited in a lake setting c. *300 million years ago, during the Carboniferous Period.*

Geological Landscape: Katherine Addison-Scott, Severnside Branch
Moereki Boulders *(overleaf, right)*
Moereki Boulders, New Zealand, in the early morning light.
[The Moeraki Boulders are concretions created by the cementation of the Paleocene mudstone of the Moeraki Formation, from which they have been exhumed by coastal erosion. They consist of mud, fine silt and clay, cemented by calcite. — Ed.]

Industrial Geology: Eileen Lawley, Mainland Europe Branch
Guilaizhuang Gold Mine, south-east China
The gold deposits date from the Middle Jurassic and the mine has been operating since 1992. The gold is extracted using a method known as Carbon Pulp and is operated in an environmentally friendly manner by retrieving the cyanide and arsenic involved in the extraction.
The average yield is 6.8 grams/tonne and the group calculated that for the given figure of 10kg to 15kg of gold extracted per day, and with the current price of gold, c. *$1,000,000 per day gross was made on site.*
[Note: A slightly lower figure is reportred by Helen Gamble on page 57. — Ed.]



OUGSME field trip to China, August 1 to 10, 2011

Elisabeth d'Eyrames, Helen Gamble, Jane Hickman, Eileen Lawley, Neil Lawley and Mike Molloy

After we had all arrived safely at various times in Qingdao, a city of seven million inhabitants in the province of Shandong, we were introduced to our Chinese guides for the trip. Chen Yuan, a student at the Beijing University of Geosciences who was to explain the geology that we were to see, and Li Wei, who had made all the travel arrangements (Fig. 1).

Figure 1 The group poses in the port area of Quindao (© Neil Lawley 2011).

Day 1 Zhu Cheng Dinosaur Geopark (2 August)

by Eileen Lawley, photos by Neil Lawley

After a short stop in the port area of Quindao, where we saw the sailing arena used for the 2008 Beijing Olympics, we drove over the 'longest bridge over the sea in the world' (24km) across the bay, saving *c.* 100km. Our main events of the day were a visit to the Cretaceous Dinosaur Geopark at Zhu Cheng and its nearby Zhu Cheng Dinosaur Museum. After passing a building stuffed with complete dinosaur skeletons, glimpsed through the windows, we approached another building, and after opening the sliding doors we entered and gasped in amazement (Fig. 2).

We could not believe what we were seeing. We were in an enormous shed 500m long and 30m wide. On one side were layers of red sandstone and mud sediments. On the other side of a pathway, stretching to the end of the shed for 500m was a 10–12m high mound formed from a strata deposited in the Late Cretaceous that was absolutely bristling with dinosaur bones — an estimated 10,000 of them waiting to be to excavated and identified by eager students. What a sight! As we walked along the gallery identifying some of the bones, we felt the impending excitement that the students will experience as they get to work on their discoveries.

The first dinosaur was excavated in 1964. There are three sites: the dinosaur fossil gallery just described, an 'uplift of dinosaur fossils' 300m × 30m and a 'dinosaur fossils stacked zone' giving a further 6,000 fossil pieces. These three sites constitute the world's largest exposed area of dinosaur fossils, so we felt very privileged to have seen one of the sites.

Ninety per cent of the fossils found so far are of Hadrosaurus (duckbilled dinosaur), which my copy of Simon and Schuster's *Encyclopedia of Dinosaurs* (1999; Marshall Editions Developments Ltd, USA) does not even acknowledge as being found in China. There are also examples of *Triceratops*, which it is claimed are the 'mother' of the species found in the USA.

Entering another shed we studied many display cases of dinosaur bones and several large assembled specimens. A couple of restorers were pleased to have a rest from scraping while they showed us their handiwork, painstakingly and lovingly executed. One bone takes several weeks to prepare, we were informed, so there is no need for any unemployment here; and they receive a fairly good wage by Chinese standards — about 160 euros/month. In comparison, you can only imagine what the costs for restoration would be if undertaken in the West.

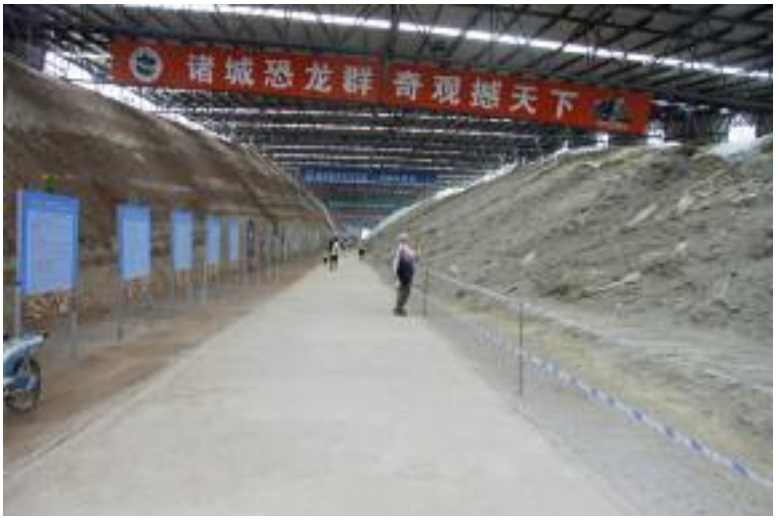


Figure 2 The dinosaur fossil gallery at the Dinosaur Geopark, Zhu Chung (© Neil Lawley 2011).

Figure 3 Gigantic Zhucheng Dragon in the museum, the world's tallest Hadrosaurus skeleton (© Neil Lawley 2011).

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The museum itself is situated in yet another building, where displays are built to be more suitable for public viewing. There were some interesting finds showing nests of up to 30 dinosaur eggs, the 'Shandong giant dragon' (standing at 8m high and 15m long) and the 'Gigantic Zhucheng Dragon', the world's tallest *Hadrosaurus* skeleton at 9.1m high and 16.6m long (Figs 3 *(previous page)* and 4).

We left rather reluctantly and drove on to our next stop at Zhucheng for the night.

Figure 4 Other exhibits in the museum with Eileen (1.5m) as a scale touching a femur for 'good luck' (© Neil Lawley 2011).

Day 2 Yingguang Lake and Guilaizhuang Gold Mine (3 August)

by Helen Gamble

The first stop on Day 2 was at Yi Shui Underground Florescent Lake, which forms part of the much larger tourist attraction of Yingguang Lake Tour. The underground lake forms part of a large karst cave system with a length of 1,200m; the lake itself covers an area of 25,000 square metres and has a capacity of 240,000 cubic metres. The natural formation is related to the unvisited 6.1km long canyon, developed from a north-west–south-east karst fracture, with both formed *c.* 65Ma to 230Ma years ago.

The cave system was an integral part of the tourist attraction, with only a few of the natural features, such as stalactites, easily visible. To gain access, we climbed aboard one of the flotilla of cockle boats that transports visitors into the main attraction. With the boatman pulling our craft along by a series of ropes attached to a timber support walls or occasionally the roof, we entered the main cave. As we were pulled along in the inky blackness, lights gradually appeared on the roof like thousands of twinkling stars in a clear night sky. These are the fireflies, a species that lives in the stable climate afforded by the cave system and their light is created by a biochemistry reaction. We explored the remainder of the park on the way back to the bus, and found a few panels translated into English, but none mentioned the geology of the area (Figs 5 and 6).

Our second stop, Guilaizhuang Gold Mine, is 20km south-west of the town of Ping Yi. We entered to the south-east of the town, and our arrival was somewhat unexpected. It was only by chance that Chen Yuan managed to talk to the site geologist, Lu Dong, who had also attended the Chinese University of Geosciences, and so were permitted access to the site.

The gold mine consists of a crypto-explosive breccia developed in Paleozoic carbonate rocks and Ordovician dolomite, and controlled by an east–west-trending listric fault with subvolcanic epithermal telluride-bearing gold deposit. The deposit is related to the Early Jurassic Tongshi magmatic complex that formed in a continental arc setting on the margin of the North China Craton. The host rocks are porphyritic and consist predominantly of medium- to fine-grained diorite and hornblende-bearing monzonite. The fault zone has the largest ore body — which pinches, swells, branches and converges — contributing 98% of the total ore reserves, with a length of 550m, an average thickness of 6.8m and a dip width of 650m (Fig. 7, *opposite*).

Our first stop on site was to an area where the rock was taken and sorted, depending on which part of the site it was extracted from. This process ensured that different grades were kept separate,

Figures 5 and 6 Cockle boats and stalactite at Yingguang Lake and karst cave (© Helen Gamble 2011).

Figure 7 (left) Ping Yi rock (© Helen Gamble 2011).

Figure 9 (right) Good Luck Stone at Guilaizhuang Gold Mine (© Helen Gamble 2011).

as the ranges of metals and minerals are not consistent throughout the site. The gold deposits range from 3.42 to 26.37 gram per ton (g/t), with a mean of 6.8 g/t and a maximum of 457.4 g/t. The group calculated that for the given figure of 10–15kg of gold extracted per day, and current price of gold, *c*. $900,000 per day gross was made on site.

The next stop was the main pit — 190m deep, with another 300m of deposits to excavate. There were also six shafts, with 40m deep adits running off every 150m. This amazing feat of engineering, and the chance to look down into it safely from the viewing platform, was truly breathtaking (Fig. 8).

Zircon-sensitive high-resolution ion microprobe (SHRIMP) U-Pb dating indicates that the dioritic rocks of the Tongshi magmatic complex give a mean age of 175.7±3.8Ma, interpreted as representing the crystallisation age of the Tongshi magmatic complex, dating the gold deposits to the Middle Jurassic.

Figure 8 Guilaizhuang Gold Mine — the Main Pit (© Helen Gamble 2011).

Operating since 1992, the site has recently undergone extensive modernisation, enabling an environmentally friendly, and more economically viable, production system to be operated. Currently the site is not officially open to the public, but it will be when all the internal roads are finished. Large karst blocks were positioned along the access track surrounding the main pit; and above the site, set in temple-like shelters, were vast stones extracted from the pit. At a cost of 200,000 Yuan, these had been erected overlooking the site operations, both for good luck and for the visitors (Fig. 9).

Lu Dong spent the afternoon taking our group around the site and explaining the five-stage process required to extract the gold and recycle the materials:

- Stage 1 Sorting, crushing and pulverising of the extracted rock to produce a fine dust material.
- Stage 2 The fine material is then added to a hydrochloric acid and sulphuric acid solution to form a mud/rock slurry (Fig. 10).
- Stage 3 The slurry solution is treated with carbon, which extracts the gold from the other metals.
- Stage 4 Sodium cyanide is then used to extract the gold from the carbon (Fig. 11, *overleaf*).
- Stage 5 The waste material is dried in settling areas and the dry material is landscaped, while the solution extracted at the end of the process is piped back down to the crushing area to be re-used in Stage 2 (Fig. 12, *overleaf*).

It was an excellent afternoon, even better than anticipated as we were taken around all the processes and machinery. The site would benefit from some form of interpretation of the processes and geology before it is opened to the public.

Figure 10 Guilaizhuang Gold Mine — processing Stage 2 (© Helen Gamble 2011).

Figure 11 (left) Guilaizhuang Gold Mine — processing Stage 4 (© Helen Gamble 2011).

Figure 12 (right) Guilaizhuang Gold Mine — processing Stage 5 (© Helen Gamble 2011).

Day 3 Qufu (4 August)

by Elisabeth d'Eyrames and Mike Molloy, photos by Elisabeth d'Eyrames and Neil Lawley

As many of us still felt jet-lagged, Day 3 was devoted to tourism. We spent the morning visiting Confucius's birthplace in the little town of Qufu (Fig. 13). Confucius lived *c.* 551 BC–479 BC, and is probably the most famous Chinese thinker and philosopher; he allegedly emphasised personal and governmental morality, correctness of social relationships, justice and sincerity. His teachings became established during the Han Dynasty, 206 BC–AD 220, some 200 years after his death. Our Chinese guide for the morning managed to complete our tour without needing to use microphones and loudspeakers, which are in general use for Chinese-language tours because of conflicting noise from simultaneous tours.

Figure 13 (left) Qufu — the birthplace of Confucius (© Elisabeth d'Eyrames).

Figure 14 (right) A view from the walkway up Mount Tai (© Elisabeth d'Eyrames).

Figure 15 (below right) Mount Tai Geopark — allgovite intrusion (© Elisabeth d'Eyrames).

We continued our journey to Tai'an (70km from Qufu), the town at the foot of the Tai Mountain, which was to be next day's destination.

Day 4 Taishan (5 August)

The aim of the day was to tackle Mount Tai (36° 16' N and 117° 6' E), 'the chief of the five Sacred Mountains', or the 'first mountain under Heaven', altitude 1,545m. It is a walkway of 6,293 paved steps, some 30–40m wide on the lower parts, but getting narrower and steeper towards the top (Fig. 14). At five in the morning we passed the first gate and started walking on a path that has been used by pilgrims and famous emperors for 3,000 years. Taoist temples have been erected along the steps and some are still in use. As the mountain is also a geopark, there are panels with geological explanations to draw the attention of the walkers to the extraordinary history of the rocks.

The rocks that make up this massif are part of the North China Craton basement rocks formed in the Archean; some outcrops are as old as 2.75Ga and are similar to the rocks of the so-called 'greenstone belt' in other parts of the world. As such they are contemporary to the rocks of north-west of Scotland that were the subject of a talk at our Stirling symposium last summer.* At 05:15 hrs, we stopped in front of a massive basic-vein outcrop, formed by a magma intrusion in the deep crust. This 1.76Ga old alkali basalt is mainly composed of plagioclase and pyroxene and named allgovite; it is blackish green on the fresh surface, and light brown on the spherical weathered surface. A little farther on, we stopped in front of a smaller allgovite vein cutting through a quartz diorite of the Zhongtianmen rock body, 2.5Ga (Fig. 15).

** See: 'The Laurentian basement of the North-west Highlands' by Kathryn Goodenough,* OUGSJ ***33*** *(1) 2012, 9.*

Figure 16 Mount Tai Geopark — the crouching lion (© Elisabeth d'Eyrames).

In Walter (1989) allgovite is mentioned as an obsolete name that referred temporarily to a group of basaltic rocks that have not been classified. It seems that it is still in use in Chinese classifications, as many Chinese texts refer to it.

At seven a.m. and 500 steps farther, we had breakfast next to an intrusion in the Taishan Group (2.55Ga). This intrusion is of quartz diorite of the Zhongtianmen Group (2.5Ga), which we had seen before with the allgovite. The Taishan Group comprises supracrustal rocks that experienced tectono-thermal events at the end of the Neoarchean. They are the most ancient strata in China and occur as lenses and strips in granitoids. Before undergoing metamorphic processes, these strata formed a 12,000m thick cover of mafic to felsic volcano-sedimentary varieties, and argilloarenaceous sediments. Recent investigations using SHRIMP U-Pb zircon dating on metasedimetary and volcanic samples indicate that the sediments of this Group were deposited in the late Neoarchean 2.55–2.525Ga. The Taishan Group has subsequently undergone greenschist to amphibolite facies metamorphism, in other words, moderate to high temperature and moderate pressure metamorphism, producing mainly leptites (fine-grained granulose metamorphic rocks), gneisses and amphibolites. In some of the literature it is said that the Group has undergone migmatisation and had formed all kinds of migmatites, partial migmatites and migmatitic granites, which is just another way to describe metamorphic processes. Recent studies conclude that a magmatic arc model might be applied to explain the tecto-thermal evolution at the end of the Neoarchean of this Group. On our way up, we had the opportunity to see more of these rocks, as the bulk of Mount Tai is composed of intrusive magmatic bodies.

After climbing another 700 steps, we had a discussion about the inscriptions on the rocks we had seen on our way up: there are some 820 stone tablets and 1,020 cliff-side inscriptions from famous scholars of different dynasties recorded on the mountain. Some of these inscriptions are so ancient that they are in Chinese characters no longer in use. Chairman Mao also climbed the mountain, and had some of the inscriptions carved.

Stone masons took advantage of the large vertical-joint surfaces of the intrusive rocks — which account for 95% of the main body of Mount Ta — to complete their work. The rocks are essentially tonalite, adamellite and diorite, and they were all emplaced during the Archean.

By nine a.m., with 1,600 steps behind us, we caught up with colleagues who had taken the bus up to the middle station. From this station, an alternative is to take the cable car the rest of the way to reach the summit. At the middle station, there is an outcrop of a beautiful onionskin weathered diorite. Such a dome-shaped pattern of weathering is due to the long-term temperature difference between day and night. The Chinese call the stone the 'lion stone', as, depending on your angle of view, you might recognise a crouching lion (Fig. 16).

The long, steep walk up to the summit begins from this middle station, so we stopped only briefly to enjoy the view and have some lunch on gneissic rocks under an ancient inscription. We reached the top of the mountain some two hours later, where we rejoined colleagues who had taken the cable car. They where sitting on a stone bench and had spent the last hour as photographic models for Chinese tourists.

On our climb we stopped at several locations where many good examples of the main rock bodies forming Mount Tai could be seen. The Yuhuanding rock body (2.55Ga) is the main one that is distributed around the summit of Mount Tai, because of its hardness and strong weathering resistance. It is a red coarse porphyritic adamellite, with large K-feldspath crystals in a K-feldspath, plagioclase, quartz and biotite matrix. Although this rock name comes from Adamello, a mountain in South Tyrol, one of the best examples in the UK is the Shap Fell Granite (as noted in the *Oxford Dictionary of Earth Sciences* (2008 edn)). It is mainly sandwiched between Aolaishan adamellite (2.4–2.5Ga), a yellowish grey medium-grained rock body, and banded gneiss of the Wangfushan Formation (2.7–2.75Ga). These last rocks, which are the oldest, relate to the beginning of the geological history of the basement. Beginning with the emplacement of basic and tonalitic magmas (2.7–2.75Ga), these were subsequently metamorphosed to amphibolites and grey gneisses shortly after their emplacement, and constitute the metamorphic basement termed the Wangfushan gneisses. The Wangfushan gneisses were later intruded by plutonic rocks such as the Yuhuangding and Aolaishan adamellite and diorites.

Figures 17 and 18 Mount Tai Geopark — relationships between the intrusive rocks (© Elisabeth d'Eyrames).

Figure 19 Mount Tai Geopark — Wangfushan gneiss inclusion in Yuhuangding adamellite (© Elisabeth d'Eyrames).

In the B. M. Jahn *et al.* (1988), their study of the gneisses and the diorites rocks associated with the Archean crustal evolution are thought to reveal that continental growth took place by the addition of juvenile material from depleted mantel sources, whereas the adamellites (granite type rocks) derive from partial melting of the basement gneisses. This model of crustal evolution has also been deduced from the study of Archean rocks in other parts of the world (Figs 17 and 18, *previous page*).

In Figures 17 and 18 (taken next to each other), the relation of the intrusion of the four rock bodies can be observed: the Wangfushan banded gneiss (1) appears as a residual inclusion in the Yuhuangding adamellite (2), which has been intruded by the Aolaishan adamellite (3), and eventually the Puzshaosi diorite (4) intruded the entire complex.

Figure 19 is another good example of the Wangfushan gneiss inclusion in Yuhuangding adamellite. The name Yuhuangding means the Jade Emperor Peak, which is the Chinese name for Mount Tai. The mountain has always attracted many people from all over China, but since 1987, when Mount Tai was officially included into both the UNESCO lists of cultural and natural heritage, together with the Great Wall and the Forbidden City in Beijing, the number of visitors has strongly increased. The positive aspect is that it is possible to get up to and down from the summit by bus and cable car, which we greatly appreciated on our way back (Fig. 20).

The age of the rocks that we saw on Day 4 reach back to the very beginning of the formation of continental crust, they have subsequently experienced a long and complicated geological history including many strong crustal movements, at least six, among them are the Caledonian event and Hercynian event, and eventually the Himalayan orogeny.

Figure 20 Close to the summit of Yuhuangding (Jade Emperor Peak), the Chinese name for Mount Tai (© Elisabeth d'Eyrames).

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Mount Tai Geopark path panels, and commentary by Chen Yuan

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Day 5 Culaishan National Forest (6 August)

by Jane Hickman

This was our last full day of geology. We left Tai An and headed south-east to Longwan Geopark in the Culaishan National Forest. Culai Mountain is said to be the sister of Mount Tai, but our visit to the area was in sharp contrast to our previous day climbing the sacred mountain. We had the place practically to ourselves, which was soothing to the soul after rubbing shoulders with half the population in Taishan, and, secondly, it was raining (Fig. 21).

Before us, heavily wooded hills were dissected by steep twisting valleys littered with huge boulders that had fallen into

Figure 21 Culaishan National Forest — a view up the valley we were to explore (© Jane Hickman 2011).

Figure 22 Culaishan National Forest — following the path up along the side of the stream (© Jane Hickman 2011.

the delightful stream, all softened by a light mist. Almost like home (Fig. 22, *opposite*).

The North China Craton comprises of three distinct blocks, of which the Eastern and Western blocks share similar geology. They are dominated by late Archean Tonalite-Trondjhemite-Granodiorite (TTG) gneiss complexes, which were regionally metamorphosed *c.* 2.5Ga. The Central Block contains Palaeoproterozoic crustal rocks as well as some Archean material that underwent regional metamorphism *c.* 1.85Ga (Fig. 23).

The North China Craton was formed by the collision of the Eastern and Western blocks *c.* 1.85Ga (Zhao 2001). Imagine the simple scenario in Paleoproterozic times where the Western Block, with a passive eastern margin is separated from the Eastern Block by an ocean. The western margin of the Eastern Block is active, and as the ocean closes the sediments of the passive margin and the island arcs of the active margin combine to form the Central Orogenic Belt *c.* 1.85Ga. Globally there were many of these collision orogens *c.* 2.0–1.8Ga according to evidence from other Archean cratons around the world.

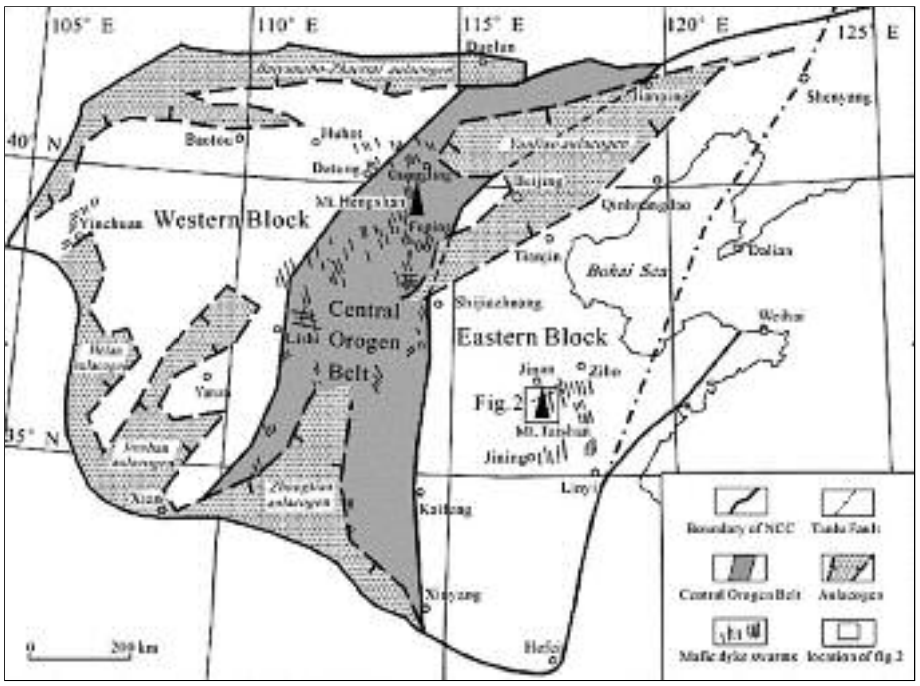
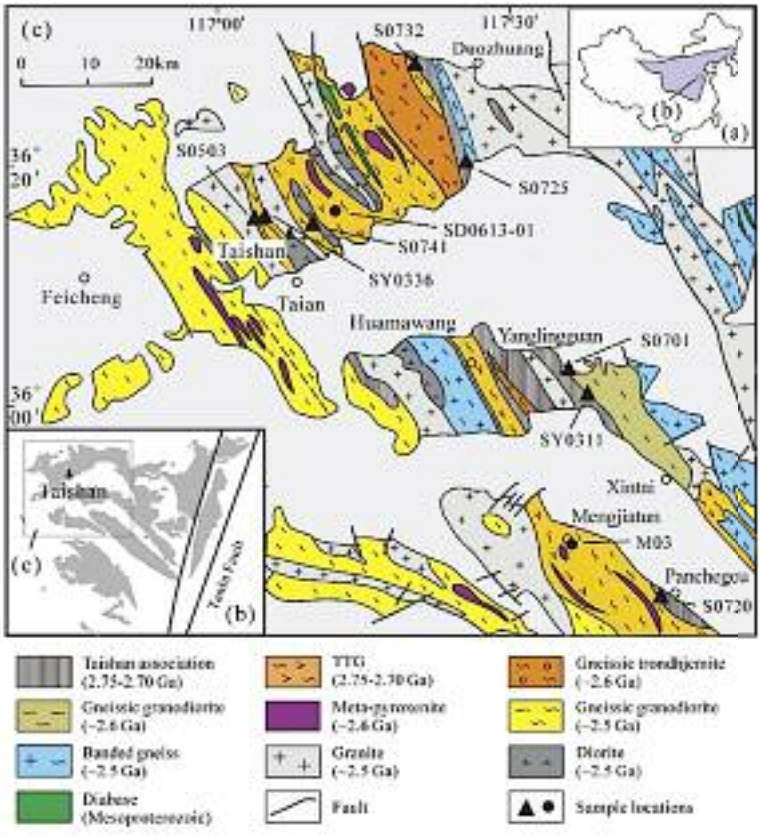


Figure 23 The three blocks of the North China Craton (G. Hou, Y. Liu and J. Li, Journal of Asian Earth Sciences, 2006).

The timing of the collision remains controversial. According to Hou *et al.* (2006) they took place between 2.5Ga and 1.8Ga. However, it is accepted that shortly after the formation of the craton there was widespread extension. Figure 23 shows the aulacogens of the rift together with the distribution of the associated mafic dyke swarms that have been dated 1837±18Ma at Taishan (Hou *et al.* 2006). The swarms throughout the North China Craton show similar dates in the range 1.84Ga to 1.75Ga, indicating that the whole area experienced continuous extension in Late Paleoproterozoic-Mesoproterozoic times (Hou *et al.* 2008). However, the extension ceased and the triple junction failed to form.

Figure 24 The Taishan Association (formerly Taishan Group) — a simplified geological map of the North China Craton (Wan et al.*, Journal of Asian Earth Sciences, 2011).*



The western Shandong Province lies within the Eastern Block and comprises a granite-greenstone belt. It broadly consists of two major geological entities: the basement gneisses and the intrusive plutonic association (Jahn *et al.* 1988).

The Taishan Association (formerly known as the Taishan Group) occurrs as strips and lenses within the gneisses and granites (Fig. 24). These supracrustal rocks underwent greenshist-amphibolite facies metamorphism and are considered to have formed between 2.75Ga and 2.70Ga (Wan *et al.* 2011).

The Taishan association is divided into three units:

1 the Yanligguan Unit, composed mainly of amphibolite, where amygdales and lava flows have been identified and typical Archaean volcano-sedimentary associations;
2 the Shancaoyn Unit, composed mainly biotite gneisses with locally preserved bedding structures and interlayered with amphibolites and mica shists;
3 the Liuhang Unit, which is lithologically similar to the Yanligguan Unit but contains more detrital materials. Recently, relic pillow lavas have been found in the Liuhang Unit (Wan *et al.* 2011).

Simplistically, the basement gneisses are amphibolite facies rocks comprising mainly of TTG. They are of similar metamorphic age to the Taishan Association and both are thought to be derived from a depleted mantle source. The granites (or adamellites) of *c.* 2.5Ga, on the other hand, are thought to be derived from the partial melting of these older basement and supracustal rocks (Wan *et al.* 2011).

Figure 25 Archean rocks of the Taishan Complex (© Neil Lawley 2011).

We had seen some of these rocks the previous day on our walk up Taishan, which is also part of the Taishan Group (or Association). However, today these rocks all looked meta-igneous to me, and I could make little sense of the geology. So I ceased to worry and just enjoyed the scenery. The path, originally a wood-gathers' track, crossed the stream in a succession of different bridges, some fixed and some swinging. The path steps were new, worked by hand from local stone, and are beautiful, especially when wet.

After lunch we continued to the end of the trail were some of us undertook the initiation ceremony for this sacred place. On our return we met up with the warden and his herd of goats.

Day 6 From Tai'an to Beijing (7August)

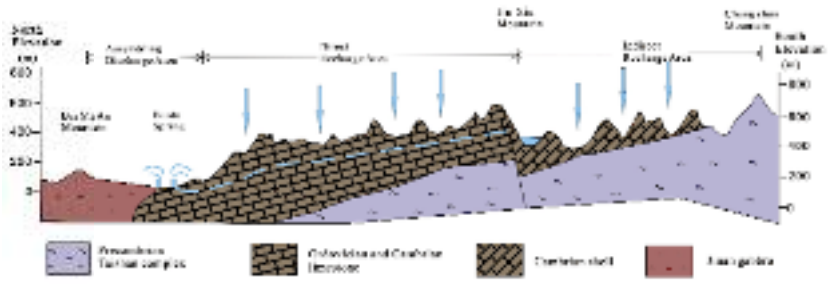
by Elisabeth d'Eyrames, photos by Neil Lawley

We left Tai'an with a last stop next to a bridge where the water erosion displayed the beautiful patterns of the Archean rocks of the Taishan Complex (Fig. 25).

Then we started the long journey north to Beijing, with a stop in Jinan — the capital of Shandong Province with six million inhabitants and a place where you can eat the best dumplings! We visited the Baotu Spring Park in the centre of the town, where there are 72 listed springs; and, of course, there are many tales about Emperors, tea and springs. The park contains several springs, among them the Baotu Spring, which is the most popular as sometimes the water gushes out when the pressure is high, although on the day of our visit we only saw the water gently bubbling in the middle of the pond. Boatu means 'leaping' and 'jumping', and many people travel from far to visit this place.

The Figure 26, drawn by Chen Yuan, reproduces the geological profile that was shown on a poster that explains why there are so many springs in Jinan. The mountain we drove through on our way from south of Jinan are covered with Cambrian shale, which allows all the rain water to flow into the limestone formation faulted against the shale. Next to Jinan the water table hits a gabbroic pluton and thus forces the water up through the karst to the surface, forming natural artesian springs. The level of the water table is constantly monitored and displayed. On the day of our visit it was at a depth of 21m.

Figure 26 The geological profile of the Jinan region (drawn by ChenYuan, 2011).



Three years earlier [than 2011], a shortage of rain during the summer had lowered the level of the groundwater to the extent that the Baotu Spring dried up, and the inhabitants of Jinan were asked to limit their water consumption. The people readily complied with this request, as they are very proud of their springs and care about them. In the park, Jane [Hickman] drew our attention towards the stromatolites that could clearly be identified in limestone blocks surrounding small lakes; they display exactly the same pattern as in the Cambrian limestone of northwest Scotland (Fig. 27).

Figure 27 Baotu Spring Park, Jinan — stromatolites in limestone blocks (© Neil Lawley 2011.

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Dinosaur footprints from the Lower Greensand *Mammilatum* Bed of Folkestone?

Philip Hadland*

** Philip has an MSc in Palaeobiology from the University of Bristol. He is the Assistant Collections Manager for Canterbury City Council and Chairman of GeoConservation Kent — Ed.*

Abstract

The Cretaceous Lower Greensand bed is well exposed on the foreshore at Folkestone. These Greensand beds are known as the Folkestone Beds and preserve a diverse ichnofauna, including footprints of terrestrial vertebrates. They reveal what appear to be three closely grouped two-toed footprints of a large vertebrate. Two of the prints are of such similar form that it cannot be denied that they are indeed footprints made by an animal. Specimens of presumed 'footcasts', some of which resemble the form of an ostrich foot, have also been found in close proximity to these footprints and collected. The footcasts are of comparable sizes to the prints and also indicate two toes. Based on the shape of the footprints it can be reasonably assumed that the animal that made them was bipedal. One can only speculate about the exact kind of animal, but it is possible that they were created by a kind of theropod. It is hoped that in the future body fossils will come to light revealing the identity of the animal.

Discovery

The initial discovery at Folkestone (Fig. 1 A) in November 2010 was of three distinct indentations in poorly cemented sands with a muddy, grey matrix enclosing the sand grains (Fig. 2, *overleaf*). What initially stood out was the striking similarity of one indentation to the next and the conclusion was quickly made that these must be footprints. Each of the footprints is *c*. 200mm across. Unusually they have two toes, and by analogy are like ratite birds that leave similar footprints. The animal was probably bipedal. Sadly, due to the soft nature of the rock in which they were exposed, the footprints have been eroding away rapidly, as is illustrated by Figure 3 *(page 65)*, and at the time of writing are unrecognisable. However, in addition to the i*n situ* footprints, specimens of similarly shaped rocks, which in all probability could not be anything other than footcasts, have also been found and collected from the immediate vicinity (Fig. 4a, b, c, *overleaf*). The footcasts are of comparable sizes to the prints and also two-toed. The sizes of the footprints suggest that the animal was of comparable height to a large ostrich, probably even larger.

Palaeoecology

The stratigraphic location is shown in Figure 5 *(page 65)*. The footcasts are presumed to be sand infills where the sandy

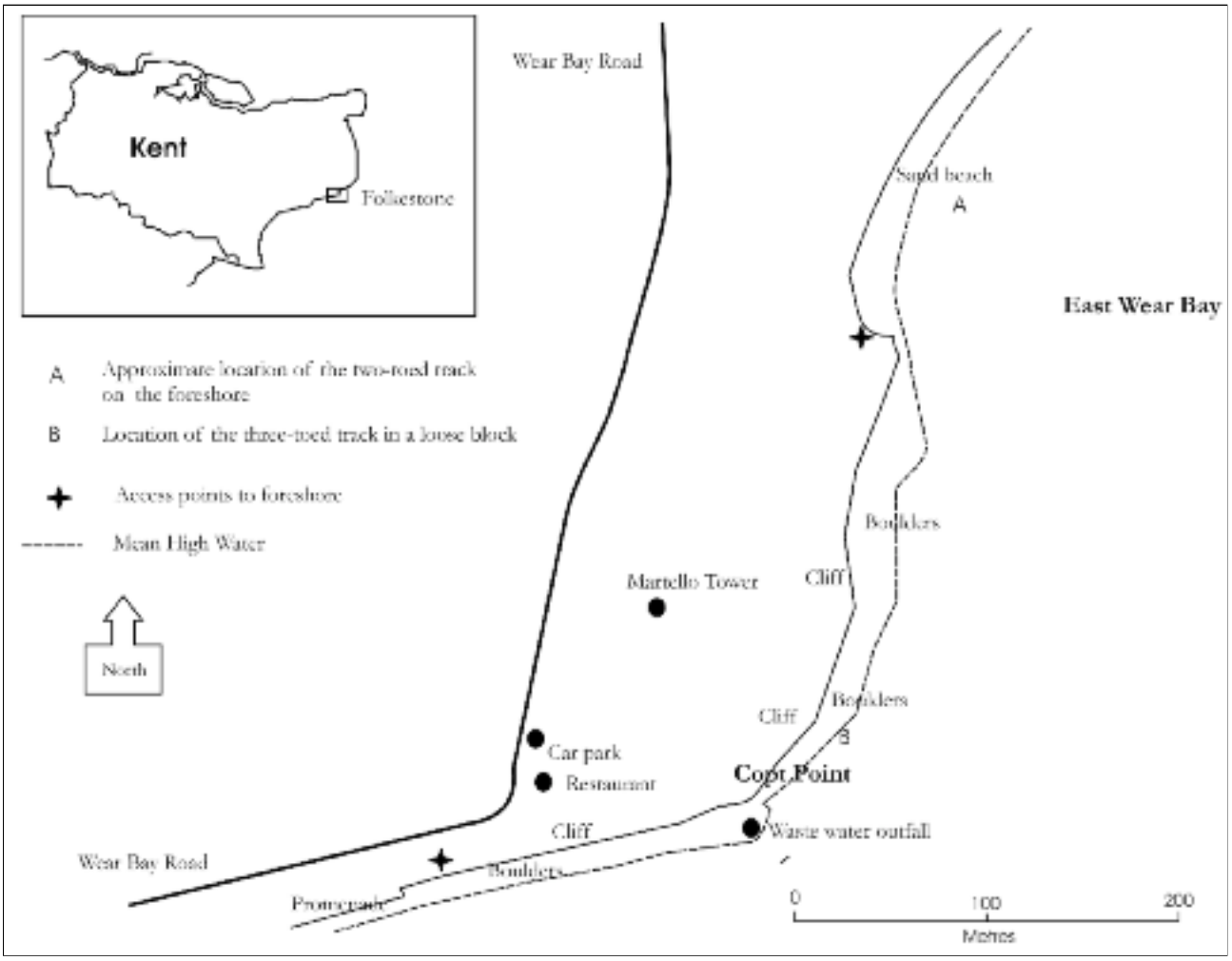

Figure 1 Map showing locations of fossil finds.

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Figure 2 In situ *footprints as first seen at Wear Bay in Folkestone. The repetition of the shape can clearly be seen.*

sediment was locally muddy. The three-toed footprint is thought to be from a layer below the *Mammilatum* Bed within the Folkestone Sand Formation. In the cliff section, the matrix of the sands above the *Mammilatum* Bed is a fine, powdery residue resembling fuller's earth both in character and behaviour.

The presence of the footprints is consistent with interpretations of the environment as shallow marine with a prograding shoreline (Anderson 1986). The presence of specific trace fossils (accumulations of balls of sand rather similar to those made by ghost crabs) in the well-cemented portions of the layer indicate tidal conditions (Fig. 6, *page 66*). The balls of sand are also found to be part of the infill of one of the footcasts (*see* Fig. 4a), which confirms their faunal association. Such crabs would have been feeding on the same abundant

Figure 4a, b, c Plan view (top) and two sets of stereo images from the undersides of footcasts collected from the same location as the footprints. Balls of sand thought to have been created by ghost crabs can be seen in the lower centre (top set: length 250mm, width 190mm; bottom set: length 200mm, width 155mm).

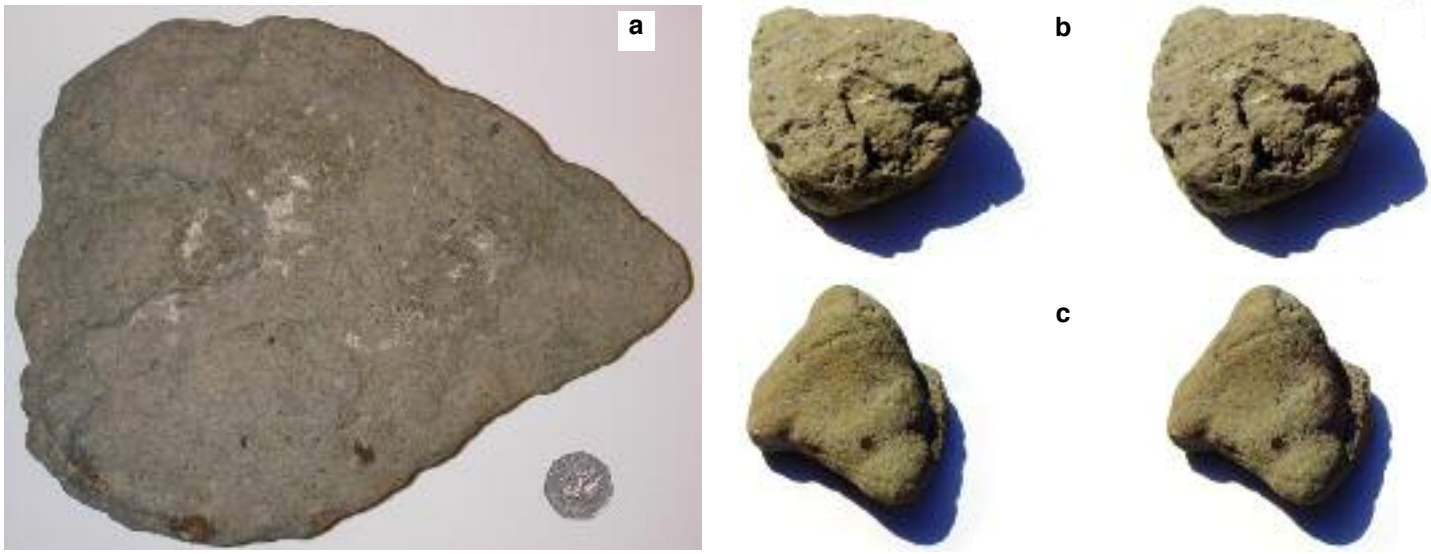

Figure 3 Photograph of footprint taken a few months later showing rapid rate of erosion (£1 coin for scale).

algae as the large oysters that are commonly found in the formation. The tidal conditions would have afforded access to the environment for large terrestrial animals. In addition, observation of poorly lithified coarse sands in the layers immediately above the poorly cemented, muddy grey matrix sands has revealed frosted quartz grains, indicating air abrasion, possibly within sand dunes (Fig. 7, *overleaf*).

A variety of reworked fossils are known from the *Mammilatum* Bed, including an abundance of ammonites such as *Douvilleiceras mammilatum*. Large oysters, brachiopods, echinoids and fragments of wood and the trace fossils are the only common fossils that can be said to have been alive contemporaneously with the animal that made the footprints. On this basis it is suggested that the animal was feeding on oysters, crabs or soft-bodied burrowing animals, much like many modern seabirds. The Folkestone Beds (of which the *Mammilatum* Bed is part) contain a diverse ichnofauna and these reveal some of the trophic structure. In particular, it seems possible that there could be a link between the footprints and conical structures identified within the Folkestone Beds (Buck and Goldring 2003). It is thought that the conical structures could be 'beak digs' made by beaked animals while feeding.

Figure 5 Graphic log, showing stratigraphic location of Mammilatum *Bed (after Owen and Gale 2010).*

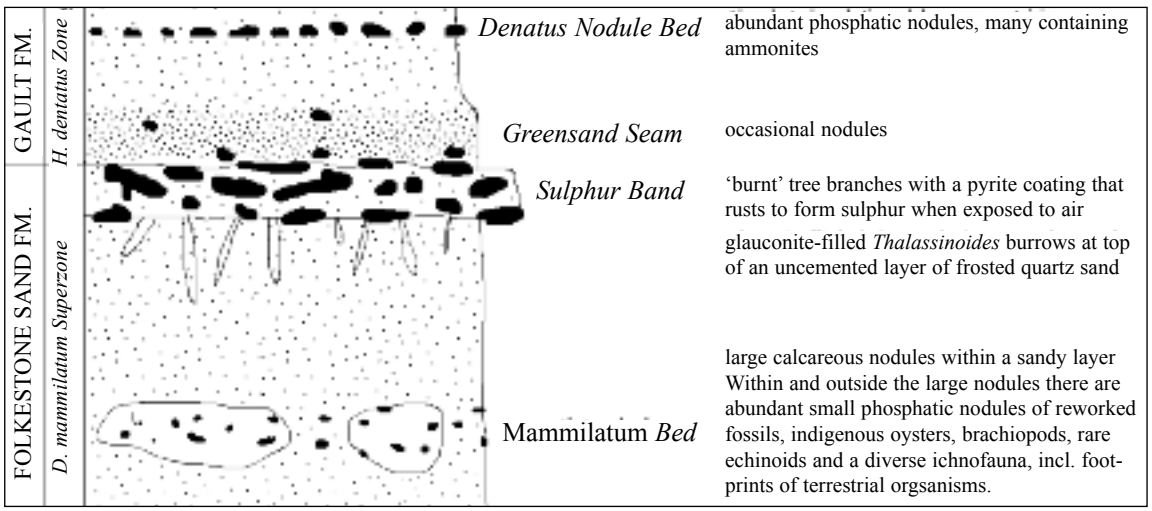
GAULT FM.
H. dentatus Zone
FOLKESTONE SAND FM.
D. mammilatum Superzone
Denatus Nodule Bed
abundant phosphatic nodules, many containing ammonites
Greensand Seam
occasional nodules
Sulphur Band
'burnt' tree branches with a pyrite coating that rusts to form sulphur when exposed to air
glauconite-filled Thalassinoides burrows at top of an uncemented layer of frosted quartz sand
Mammilatum Bed
large calcareous nodules within a sandy layer
Within and outside the large nodules there are abundant small phosphatic nodules of reworked fossils, indigenous oysters, brachiopods, rare echinoids and a diverse ichnofauna, incl. footprints of terrestrial orgsanisms.

Figure 6 (left) Accumulations of 'sand balls' interpreted to have been made by a kind of crab similar to modern ghost crabs.

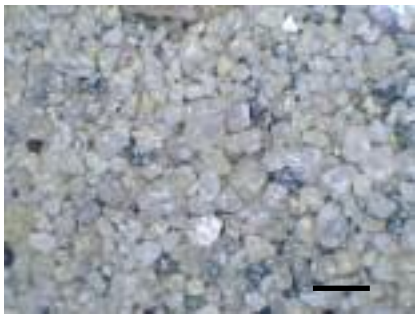

Figure 7 Quartz sand washed and sieved from the upper Folkestone Beds, just below the Gault Clay. The frosted surfaces of the grains indicate these were subject to air abrasion (scale bar is 1mm).

Conclusion

I suggest that the footprints and footcasts were made by a giant flightless bird. The recent discovery of a giant bird jaw is also interesting in this context (Naish *et al.* 2011). Three-toed dinosaur footprints are also known from the Lower Greensand at Folkestone (Fig. 8; *and see* Fig. 1 B) so the discovery of these new prints hopefully adds weight to the theory that giant birds once roamed the land alongside dinosaurs and extends their time period. It is hoped that bones will be discovered from the same formation in the near future, revealing the likely identity of the kind of animal that made the footprints.

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Aknowledgments

Philip Manning for his help with identification of the initial find and David Anderson for the 3-D photography of the footcasts.

*Figure 8 Tridactyl dinosaur footprint in loose block of Greensand from Copt Point, Folkestone (*see *Fig. 1 B). The print is* c. *120mm across and* c. *100mm in length. It is possibly a small ornithopod footprint but a theropod maker cannot be ruled out.*

Caledonides of Arctic Norway and Sweden

The Suture on a Shoestring with Severnside Branch in Lapland, Troms, Nordland, Vesterålen and Lofoten

(Led by Dr Mark Anderson, The University of Plymouth)

Linda Fowler, with Tamara Beck-Slinn, Richard Buxton, Janet Hiscott, Fran Kirby, Barbara Norton, Colin Rownes, Paul Speak, Gill Toney and Susan White

Introduction

In 2002 Mark Anderson, rather against his better judgement — as the area was completely new to him — had shown us the Norwegian Caledonides in southern Norway. That was an excellent trip but he had always said that he could do better justice to this orogenic belt if we went to northern Norway and Sweden, where he could include his PhD field area around Abisko, an area he knew much better. This trip, organised by Jan Ashton-Jones for Severnside Branch, was the result.

Mark and some of his students were already there; several of us flew from Gatwick airport; and three more joined us in Oslo for the flight to Harstad/Narvik airport. From here we drove hire cars to Abisko, just over the border in Sweden's Lapland district. With a stop in Bjerkvik for supplies, we headed to our first base, a hostel above the railway line that runs from the Swedish Baltic Sea coast at Lulea, through Kiruna — which we would visit — and on to the ice-free port of Narvik on Norway's west coast.

The field area lay between 68°–69° N and 13°–20° E: an upward curving arc comprising the Lofoten Islands and Vesterålen in the west through the north–south Caledonide chain, and the area around Lake Torneträsk to the iron mining town of Kiruna on the crystalline basement of the Baltic Shield in the east.

Day 1 Kiruna Mine visit (26 July)

by Linda Fowler

After a briefing by Mark (who cleverly projected his PowerPoint presentation onto someone's sheet sleeping bag hung over the fridge door!) we headed south-east along the southern shore of 65km long Lake Torneträsk, leaving the Caledonian belt and crossing onto crystalline basement rocks of the Svecokarelian orogen (1.96–1.75Ga), and arrived at Kiruna, where we were due at Luossavaara-Kiirunavaara AB's (LKAB's) mine at 1pm. In the local Sami language *luossa* = salmon, *kiiruna* = ptarmigan and *vaara* = mountain. After kitting out, we were driven by minibus into the mine to the 540m level (0m is the mountain summit and 230m the outside ground level where we entered). We began to realise the scale of this, the world's largest, most modern underground iron mine, as we drove deep into the mine down roads big enough for lorries and furnished with road signs, to the 'Info-Mine' area. Here we were shown an introductory film in a theatre with rock-bolted ceiling and spraycreted walls before being shown exhibits and displays, and told something of the history and geology of the deposit as well as the processing (Fig. 1).

The origin of the ore body is enigmatic: it is a 4km long sheet-like body, estimated to be 2km or more deep, dominantly of magnetite, but with small amounts of haematite, 80–100m thick, striking north–south, dipping 65°E and plunging 45°N. The hanging wall is rhyodacitic ignimbrite and the footwall a volcanic

Figure 1 Underground at LKAB's Kiruna Mine. The display shows a section through the mine and the ore body (darker grey) (© Linda M. H. Fowler 2011).

pile of trachyandesite porphyry; contacts are sharp and sheared. Its age is *c.* 1.88–1.89Ga. We were told that low temperature and pressure conditions with a calcalkaline volcanic association produce the magnetite and apatite 'Kiruna type' ore. Conflicting theories for its origin suggest that it is either of magmatic or of hydrothermal origin.

The present main level is at 1,045m with a deeper main level at 1,365m planned for 2013. The mine is worked from the footwall and the hanging wall is allowed to collapse as an area is worked out, a method known as sub-level caving. Problematically Kiruna city is built over the hanging wall, and as mining moves deeper and deeper the workings are moving closer and closer to the city, which will eventually have to be relocated, most likely to the north-east, as much of the other land in the area is Sami reserve, which cannot be developed.

Because of the shape of the Kiruna ore body it is particularly straightforward, and therefore very economical, to work. Compare this with LKAB's nearby Malmberget Mine, which is made up of *c*. 20 discreet ore bodies of various sizes, steeply dipping and not horizontally extensive. At Kiruna, after drilling and blasting, the ore is shovelled out, taken to a central crushing plant and then to the surface works where it is processed into small greenish coloured pellets — we got to know the pellets well, as they are loaded onto trains of open trucks — each train 750m long and carrying 9,000 tonnes. Ten of these trains a day carry pellets to Narvik for shipping, and all along the railway there is a liberal scattering of them! During processing the ore is separated by magnetic separators, mixed with binders and additives such as olivine, quartzite, limestone and dolomite, fed into drums and rolled into small balls. The processing uses 2% of the electricity produced in Sweden.

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Figure 2 Kiruna Mine at the surface (© Linda M. H. Fowler 2011).

Following the underground visit the minibus took us up to the top of Kiruna mountain: from here we could see the extent of the surface works above the present-day underground mine (Fig. 2), and also an exposure of the underlying volcanics in a small surface working (Fig. 3).

Returning along Lake Torneträsk we stopped to look at topography influenced by the Parvie Fault (Fig. 4). *Parvie*, another Sami word, means, appropriately, 'wave in the ground'. This significant fault is post-glacial, >9000 years old, and shows in the landscape as a 10m high scarp running 150km across northern Sweden and extending to 3km in depth. Two mechanisms have been invoked: isostatic rebound, or neotectonic stresses produced by compression of the Norwegian margin by Atlantic spreading.

Figure 3 Investigating an old working on the mountain top. Kiruna city in the background is sited above the hanging wall of ore body in the underground mine (© Linda M. H. Fowler 2011).

It has even been suggested that this could represent the first signs of subduction on the Atlantic seaboard.

Finally we stopped at Steinbeck (Fig. 5, *opposite*). Mark told us that we had moved back up from the crystalline basement into a horizontal Caledonian succession, which can be seen as alternating beds of sandstone and shale in the scarp faces.

We discussed the mechanism for the thrust sheet: it can't be a gravity driven slide since it dips back to the west (although that isn't apparent here). Was it pushed? It isn't easy to overcome friction and likely that the rocks would fail. The most likely explanation is that this is a wedge pushed in between the lower rocks and those originally above this but now eroded. This Luopokte location is the type section through the Cambrian in this area.

Day 2 Abisko National Park (27 July)

by Gill Toney

We began our overview of the thrusts and units of the early Caledonian Orogeny (490–470Ma) in northern Sweden at a roadside

Figure 4 The Parvie Fault on the far side of Lake Torneträsk (© Linda M. H. Fowler 2011).

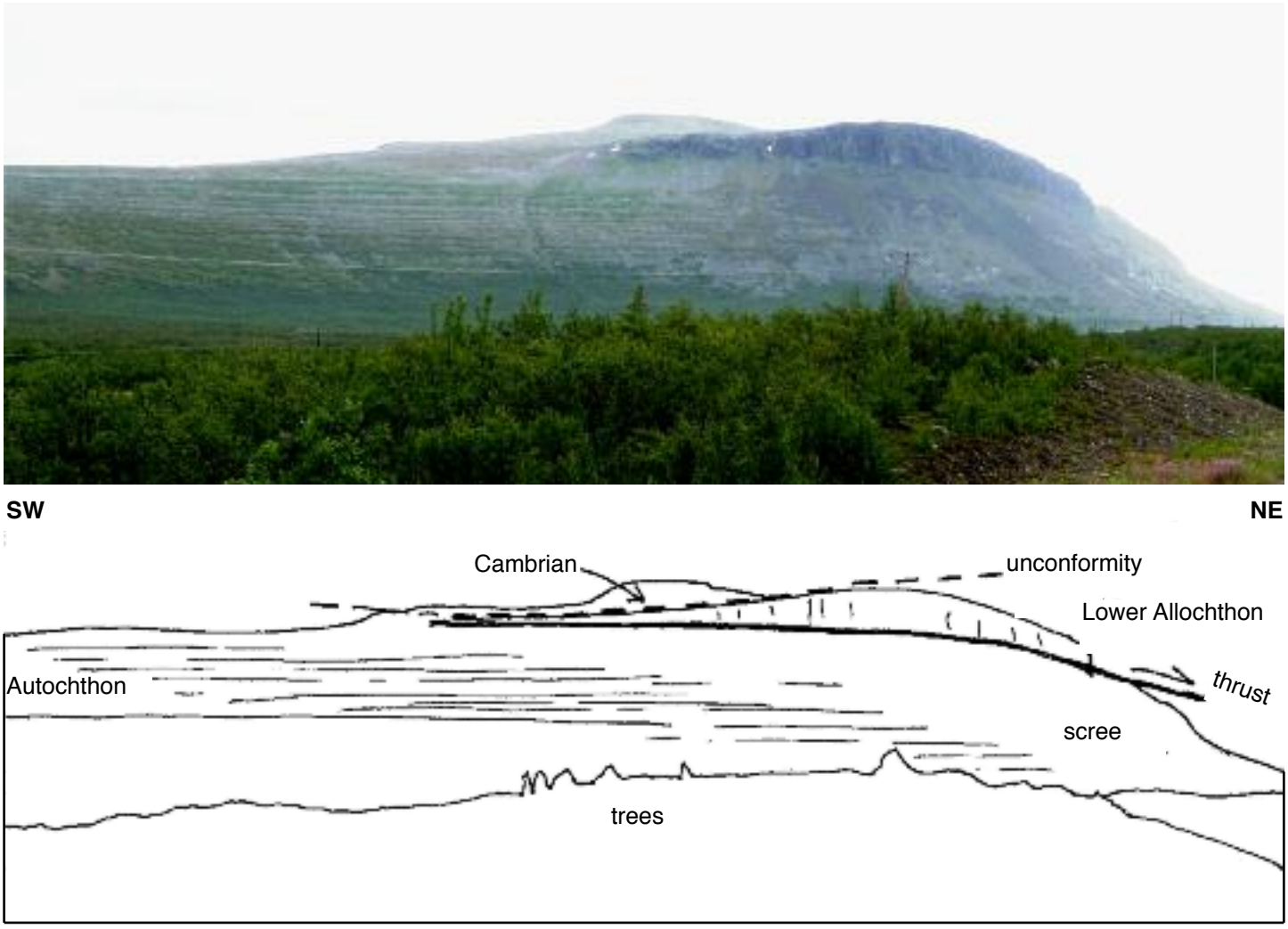


Figure 5 Steinbeck. There are alternating beds of sandstone and shale in the scarp slope; these are part of the autochthon. The vertical cliff at the top right is formed by a thrust sheet of Precambrian rocks — the Lower Allochthon — which has been transported east. Above this again there is an unconformity below Cambrian rocks of higher metamorphic grade, which can just be seen on the skyline (© Linda M. H. Fowler 2011).

exposure south-east of Abisko, where we could observe the eastern edge of the Caledonian front. Before erosion this front was originally 100km farther east. The exposure consists of deformed Cambrian metasediments, originally mudstone and clay, and is part of the repeated cover succession of the Lower Allochthon, above the Pre-Cambrian basement, which was transported from west-north-west to east-south-east over the Baltic Shield as part of the Lower Thrust Complex. There is evidence of deformation in the form of streaking in the mudstones. Fossil evidence is rare and any fabric seen in the metasediments could be a reinforcement of the original fabric. Laboratory work has shown that the unit underwent epizone metamorphism at about 300°C, on the border of the greenschist facies, indicating it was once part of a subduction zone.

The second stop involved a walk up the Abisko river canyon in the Abisko National Park to examine exposures of the Middle Allochthon. Like the Lower Allochthon, the Middle Allochthon was originally part of the continental margin of Baltica. The unit consists of a hardschist rock type known as ultramylonite, which contains quartz and is feldspar rich. There was clear evidence of intense shearing during the thrusting event, with a strong stretching lineation (Fig. 6) caused by changes to the crystal lattice during intense stress and shearing at temperatures of 350°–400°C (biotite grade metamorphism). These processes also resulted in ellipsoids lengthened in the direction of shearing (WNW–ESE).

The reduction in grain size and banding seen in the exposure are both the result of the deformation and were formed by remineralisation of the original rock. The intense west-north-west to east-north-east lineation is fairly consistent over a wide area. Isoclinal recumbent folds associated with the deformation are parallel to the lineation, as are the hinge lines together with their

Figure 6 Ultramylonitisation results in a strong stretching lineation striking at 105° (© Gill Toney 2011).

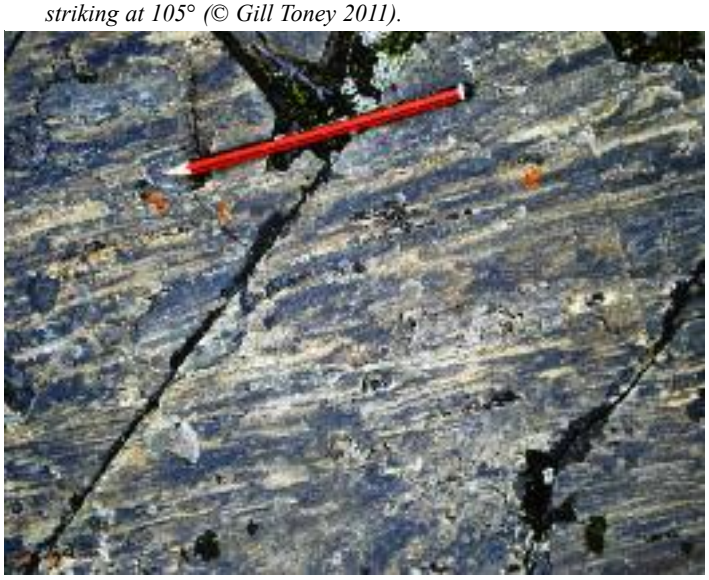


Figure 7 Isoclinal folds seen on the vertical surface have hinges parallel to stretching lineations seen on the lower, sub-horizontal surface; Abisko river canyon (© Linda M. H. Fowler 2011).

axial planes (Fig. 7). Both the folding, which thickens the sequence, and the stretching, which thins it, are consequences of the thrusting. A layer of dolomite sits in the ultramylonite unit, evidence of changes in sea level before the units were subducted (Fig. 8).

Retracing our steps we crossed the river and walked up the hillside beside a tumbling stream to the bottom of a waterfall — a vertical distance of 580m. The walk took us up through the hard-schist ultramylonitised rocks of the Middle Allochthon to its junction with the Upper Allochthon, where we expected to find a higher-grade metamorphism. The base of the Upper Allochthon is, however, a grey marble with a saccharoidal (sugary) texture. Above this is a garnet-mica schist unit, as predicted. The contact between these two units represents the suture between oceanic and continental crust.

After lunch, we took the chairlift up Mount Nualia from where we could view the Seve Nappe across Lake Torneträsk. Here the Seve is part of the Middle Allochthon and comprises a series of metasandstones with dykes, which were intruded on the outermost stretch part of Baltica and date between 490Ma and 470Ma. The unit thins dramatically across the lake and is not exposed on this side. The Køli Nappe sits on top of the Seve, forming the top unit of Mount Nualia and also the lower part of the Upper Allochthon. The grey marble unit is exposed at the base and contains amphiboles. Above is the garnet-mica schist, which contains iron-rich garnets, some with strain shadows parallel to the lineation. We also saw dark biotite and ilmenite needles.

Our final stop of the day was a roadside cutting (interestingly in a marked 'avalanche zone', though with zero snow up above we felt fairly safe) containing the upper part of the Seve unit, which here is the base of the Upper Allochthon and formed of quartzite psammite. Evidence of Caledonian deformation can be seen in the form of S-folds with hinge folds parallel to the direction of transport (east–west) and could be related to the structure that controls Lake Torneträsk (Fig. 9). At the top of the exposure sits the next unit of the Upper Allochthon, which is as yet unmapped, but shows evidence of a sharp brittle thrust that must have occurred late in the thrust event.

Figure 8 Buff-coloured dolomite unit (0.5m thick) between units of grey, ultramylonitised rock; Abisko river canyon (© Gill Toney 2011).

Day 3 Abisko National Park (28 July)

by Richard Buxton

We spent the third day a little farther west in the area between Abisko and the national border at Riksgränsen, where we examined styles of thrusting in the Lower and Middle allochthons and the tectonic window (inlier) at Kuokkel.

We first examined the Middle Allochthon at a road cutting close to the visitors' car park at Tornehamn, at the north-west end of Lake Torneträsk. Here, the rock was more pelitic than the hard-schist we had seen in localities farther east, indicating a deepening of the shore off Baltica; the rock was more brittle as we went west, due to the proximity of a major thrust. We saw good examples of

Figure 9 Roadside exposure of S-folds in the Seve Nappe — Abisko National Park (© Gill Toney 2011).

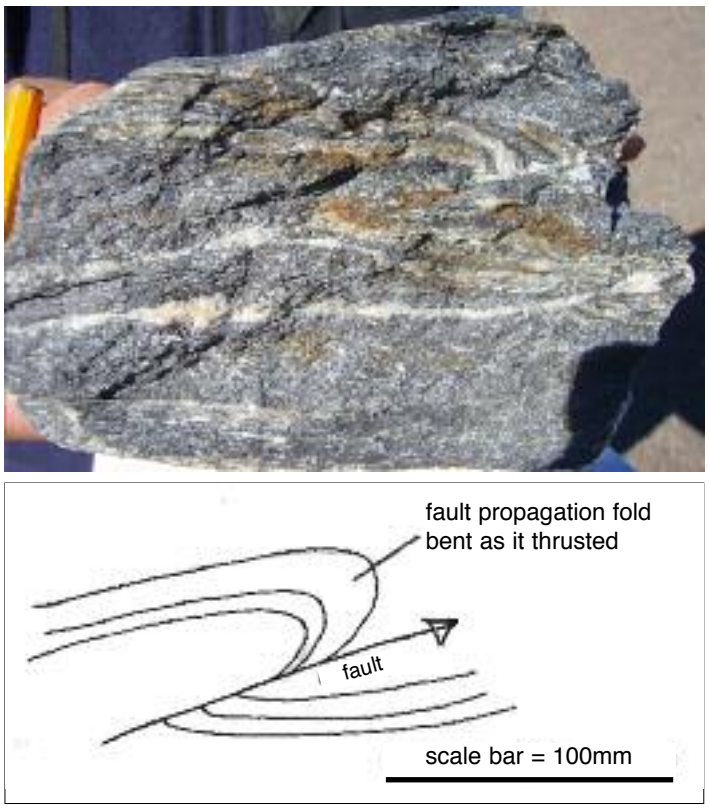


Figure 10 Fault propagation in brittle pelitic rock of Middle Allochthon — Tornehamn (© Richard Buxton 2011).

fault propagation folds (Fig. 10); and *Schistosité–cisaillement* (S–C) structures in sections parallel to the transport direction, which develop preferentially in pelitic rocks and indicate intense shearing. The sense of shear (i.e. the 'flow' of the thrust) was confirmed as being east-south-east (Fig. 11).

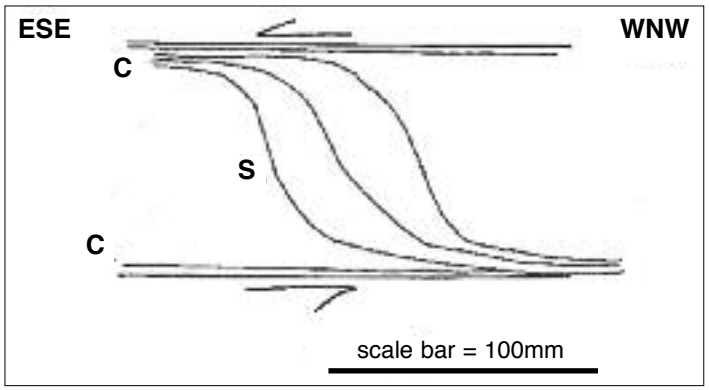


Figure 11 S-C structures show sense of shear to east-south-east — Tornehamn (© Richard Buxton 2011).

The edge of the thrust sheet has been rotated so that it is parallel to the direction of travel and is manifest as lateral shearing east-south-easterly along Lake Torneträsk. This zone of deformation is almost a fault running up the lake, and was observed in the crystalline Precambrian basement of the Middle Allocthon at a cliff site on the edge of Lake Torneträsk. Here we saw highly sheared granites trending west-north-west to east-south-east, which had metamorphosed in the biotite zone during the Caledonian orogeny. In parts, the biotite, feldspar and quartz were stretched out to form mylonite.

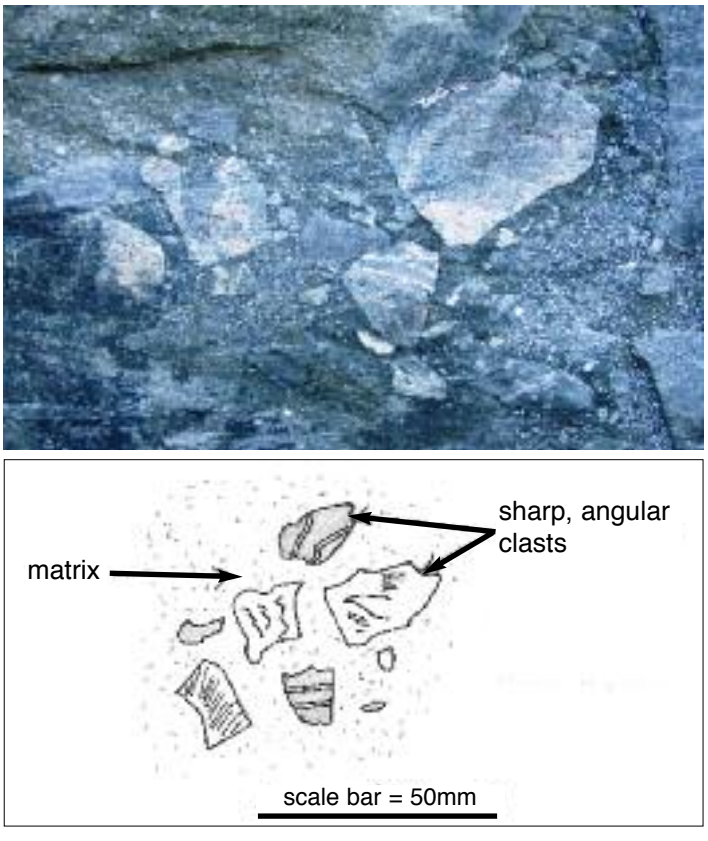
At a beach locality farther east we observed a small synform, indicating a strong 'wrench' trending east-south-east to west-north-west along the lake. Here we observed phyllitic marble of the Upper Allochthon in the central core and inward-dipping pelitic strata of the Middle Allocthon on either side.

We examined the Kuokkel tectonic window (inlier) at two sites north-west of Lake Torneträsk. At the first site, the lowest part of the cover of Lower Allocthon to the basement was found to consist of metamorphosed matrix-supported conglomerate (Fig. 12).

This conglomerate could be interpreted either as a diamictite (glacial till) or possibly a fluvial deposit, as there is some cross-lamination and scouring in it. The second exposure was in a road cutting farther west, where we examined some augen gneiss of the Lower Allocthon. We found big potassium feldspars arising in coarse-grained granites that retained their Precambrian north–south fabric, practically unaffected by Caledonian tectonics (Fig. 13, *overleaf*). The whole rigid block had translated eastwards on a brittle thrust beneath it, retaining its existing fabric and not undergoing much internal deformation.

Lastly, we travelled farther west, then walked east from Vassijaure station to view the boulder field in the Karkevagge Valley, a rock fall that occurred after the last glacial period. It consisted of huge truck-sized boulders filling a U-shaped hanging valley, oriented west-north-west, with ridges trending east-south-east–west-north-west. The boulders could be interpreted either as a massive glacial deposit, the ridges representing glacial moraine; or as a major rock avalanche arising from rock failure. A huge scar was observed in the mountain behind, which would fit with either hypothesis. Either way, the event was relatively recent and has implications for assessing risk of possible future rock falls or denudations of the land surface. *(LF note: unfortunately no pictures were taken, but this scar shows as a 2km long boomerang-shaped feature on Google Earth 34 W 389188.92m W, 7590612.97m N).*

Figure 12 Matrix-supported metaconglomerate of Lower Allochthon in the Kuokkel window (© Richard Buxton 2011).



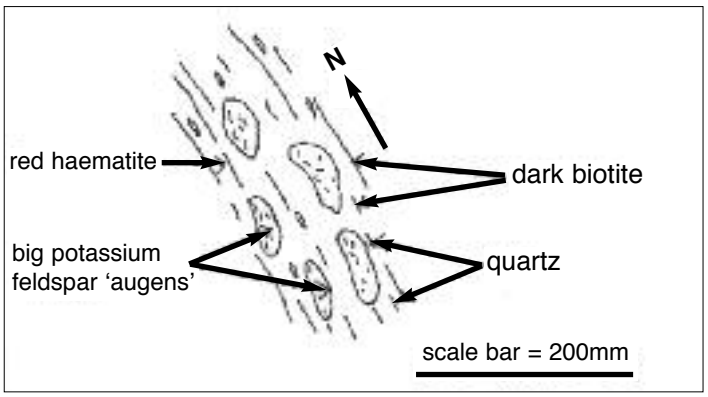


Figure 13 Precambrian augengneiss at the Kuokkel window —the pre-existing north–south fabric is maintained (© Richard Buxton 2011).

Day 4 Abisko to Lavangen (29 July)

by Janet Hiscott

We left the hostel in Abisko and headed back to Norway. En route we examined the Rombak/Sjangel tectonic windows. The exposures here are of the much older basement rocks on the Baltica margin. The allochthons that we had been looking at all week went over the top and ended up in Sweden. Late-stage uplift in the Tertiary led to erosion, which exposed the underlying basement. It is now ice-scoured classic '*roche moutonée*' terrain.

At the first stop we were on a coherent block of pre-Cambrian basement — augen gneiss. There was a steep fabric but no Caledonian influence visible. There was some jointing but this is related to the Tertiary uplift. One of the puzzles is how far the heat transferred downwards from the allochthons into the basement. Here we could see that it had had little effect; however, 400m farther along the road a low-angle fabric is present, the feldspars were much more fractured and the quartz and biotite had recrystallised and were much finer grained. This indicated that the rock had been heated to between 350°C and 500°C. Another 400m farther along and we were looking at mylonitised granite, sheared and much more aligned with the direction of Caledonian transport. Biotite ages from here give 470Ma for the fabric, which was an unexpected result as previously there had been no known event at this time affecting the Baltica margin. Kinematic indicators show that the basement was rotating into a shear zone in line with the allochthon. The basement must have been active (i.e. underthrusting causing doming) as the allochthons came over the top.

Still another 400m onwards and we were looking at the Middle Allochthon. Here the back folds in the quartz veins are interpreted as a back-sliding event to the west, caused as the body of rock moved up a significant ramp before flattening out again, i.e. the allochthon being pushed 'uphill' over the basement, but crumpling backwards when the pressure relaxed slightly.

Near by, an open fold exists with pre-Cambrian quartzite at the bottom and phyllonite (mylonite with some reconstitution of mineral grains) on top; the fold trends north–south, the existing trend in the basement. The phyllonite has a strong fabric dipping into the quartzite (Fig. 14). In the other direction we walked over ice-scoured bedrock as we walked toward a small lake

Figure 14 Open fold: the roadside exposure near Skitdalsvatnet is about 6m high, with pale grey quartzite forming the bottom half, a brown-grey, thinly bedded rock over that then slightly greyer phyllonite (© Janet Hiscott 2011).

Figure 15 Memorial to Yugoslavian prisoners at Skitdalsvatnet (© Janet Hiscott 2011).

Figure 16 (below) Thrust slices exposed at Fossbakken (© Janet Hiscott 2011).

(Skitdalsvatnet). The Cambrian conglomerate cover here strikes at 90° to the direction of Caledonian transport.

During a brief historical interlude we viewed a memorial to 4,200 Yugoslavian prisoners held in a nearby camp during WWII. They were civilians participating in the partisan movement and so, as they were not considered prisoners of war, they were not protected by the Geneva Convention (Fig. 15).

Leaving Sweden and Baltica behind, we drove back into Norway, en route to Fjellkysten ('Mountain Coast') Hotel, our traditionally grass-roofed destination for the next three nights. At Fossbakken, we saw evidence of the underthrusting to the basement. We examined the amphibolites on the roadside and identified imbricate slices with pre-Cambrian fabric. Various thrust slices are clearly evident (Fig. 16).

A key lesson for me from today was that these nappes, such huge bodies of rock, do not move as one coherent block, but rather in small packets, almost edging along. This explains why shear zones can be discrete pockets of deformation in the middle of otherwise coherent strata.

Day 5 Gratangen section (30 July)

by Colin Rownes

We spent the day studying the stratigraphy of the Upper Allochthon, looking at the stack of nappes exposed in rocks of the Ofoten synform along the Gratangen section, mainly seen in road cuttings around the end, and along the south side, of Gratangenfjord.

We started in a small quarry and then moved on to a road cutting to look at the lowest unit, the Høgtind Nappe. This is a unit of thick marble sequences inter-bedded with pelitic garnet-mica schists; quartz veining shows boudinage structures (Fig. 17, *overleaf*). Close to the boundary with rocks of the Middle Allochthon the garnets have retrogressed: they are partly replaced with chlorite, and are sheared out, so are much less common. The expression of the boundary between the two units is not clear, but can be inferred from a jump in metamorphic grade, into the garnet zone, and intense shear fabrics in the rock.

Location two, still in the Høgtind Nappe and another road cutting, was at the base of a little antiform reflecting imbrication structures at depth in the underlying basement. Rocks in the

Figure 17 Folding and boudinage of quartz veins in the Høgtind Nappe — south shore of Gratangenfjord (© Linda M. H. Fowler 2011).

outcrop show small-scale extensional crenulation cleavage (Fig. 18). This cleavage could be a problem, as we are here in a crustal shortening, compressional stress regime. It was postulated that these extensional features developed when rocks of the thrust sheet forming the nappe passed up a ramp in the basement as thrusting continued.

The second unit of the Upper Allochthon, the Grønfjellet Nappe, outcrops in the hillside above the second location. We did not visit it, but consists of garnet-mica-schist dominated amphibolite sequences metamorphosed into the staurolite zone, a higher metamorphic grade than the rocks of the Høgtind Nappe beneath.

We looked at the third unit, the Kvernmo Nappe, at in roadside cuttings along the Gratangsbotn road near Storfossen waterfall. These rocks are high-grade gneiss with plagioclase-rich granite lenses. Minerals present include coarse muscovite, biotite, garnet, kyanite, feldspar and quartz; sillimanite has been found in thin sections of these rocks but the crystals are too small to be seen in hand specimens. Again, we have here an increase in metamorphic grade, compared with the rock units below it, to the kyanite-sillimanite zone. The fourth unit, the Øse Nappe, does not outcrop in this section so we did not see it.

However, we saw the fifth unit of the Upper Allochthon at location four, another roadside section, where we looked at rocks that are at the top of the Småtinden Nappe. This is a suite of inter-bedded marbles and garnetiferous schists. The garnets in this part of the section are of particular interest, being 6–9mm in size and having ideoblastic shape, with no strain shadows evident around them. These garnets are post-tectonic and grew after movement had finished (Fig. 19). Mark posed the question of what heat source enabled this mineral growth; a possible answer was developed later. Minerals present are garnet, biotite, muscovite and quartz; we have now dropped down metamorphic grade back into the garnet zone. In this nappe there is a very useful marker horizon, the beds of the Bø Quartzite, which are a prominent series of inter-bedded quartzites and pelitic schists that outcrop over a large area. These rocks, observed in outcrop across the fjord, had a yellow-grey colour due to weathering.

Figure18 Small-scale crenulation cleavage at the second stop — Høgtind Nappe, Gratangenfjord (© Colin Rownes 2011).

Figure 19 Examples of 7mm garnet porphyroblasts at Location 4 — these grew post-tectonically over the metamorphic fabric and have no strain shadows (© Linda M. H. Fowler 2011).

At location five, another roadside exposure, we looked at beds of the Gratangen Nappe, the next unit up in the nappe stack. The rocks of this unit are grey, garnetiferous schists, with lots of quartz eyes, and inter-bedded with thin marble horizons. The garnets present show strain shadows and envelopes around them, implying that deformation continued while the rocks were still hot. The units contain the minerals biotite, muscovite, garnet, quartz, feldspar and kyanite, indicating an increase in metamorphic grade in the Gratangen Nappe to the kyanite zone. One possible source of heat for the post tectonic garnets in the Småtinden Nappe could be if the hot Gratangen Nappe was thrust over the Småtinden Nappe, providing a source of heat for garnet growth in the underlying rocks as the nappe stack cooled.

At the final location of the day we looked at rocks of the Niingen Nappe, although there was some debate as to whether we might still be in the upper units of the Gratangen Nappe at this location. The rocks seen are garnet-mica schists inter-bedded with marbles and amphibolites. These units have granite boudins present on a scale of 0.5m to 10m in size (Fig. 20, *opposite*). Rocks of the Niingen Nappe, visible in the hill tops on the far side of the fiord, consist of locally migmatised granite, schists and gneisses. This localised partial melting indicates a further increase in metamorphic grade and increased depth of burial compared with the rocks lower in the nappe stack. The presence of some kyanite, although not abundant or obvious, indicates that we are again in the kyanite-sillimanite metamorphic zone.

In summary we saw a double, inverted metamorphic sequence of rocks in this section and it can be inferred from this that a significant boundary must occur between the two metamorphic sequences: the Øse Thrust. To finish for the day we drove back to the hotel around the peninsula and down the south side of Lavangenfjord to Tennevoll, admiring the views as we went.

Figure 20 Large boudin in the Niingen Nappe — south side of Gratangenfjord near Tjuvskjer (© Linda M. H. Fowler 2011).

Day 6 Spansdalen to Hakavika (31 July)

by Paul Speak

Another hot sunny day, we returned to Tennevoll and walked up a track, off which Mark turned into the forest and, after a short search, led us to a small dam. It was so overgrown we could barely see Mark on the exposure. The rocks are kyanite-bearing migmatitic gneiss with a foliation dipping north-west. This is the top of the Kvernmo Nappe. At yesterday's location the nappe was hundreds of metres thick, but here it is only a few metres thick and does not appear at all farther north. It has been cut out by a major fault: the Øse Thrust.

About 1km north we found ourselves in the Øse Nappe with low-grade (chlorite-biotite) schists. The lower unit is dominated by mafic igneous rocks showing much shearing and stretching. The upper unit is biotite-rich schists and marbles. This sequence is thought to represent an ophiolite *c*. 10m thick. Here it is unrecognisable as ocean crust but farther north, near Tromsø, it becomes more obvious. However, Mark expressed some scepticism.

Back at the south shore of Lavangenfjord we set off to walk north along the beach, which was littered with filleted salmon carcasses. A small area of highly sheared and stretched greenish rocks was again interpreted as the ophiolite, the base unit of the Øse Nappe. Immediately above is a matrix supported conglomerate, which is the upper unit of the Øse Nappe. Higher in the upper unit are marbles and schists, intensely sheared and showing evidence of post-obduction metamorphism. The ophiolite has been dated at *c.* 470Ma and the metamorphism *c.* 400Ma. This unit would have been at the bottom of the nappe stack before obduction.

After lunch we returned north to Sagfjord and set off along the south shore as the tide started rising. We found ourselves in the Høgtind Nappe, the lowest unit of the Upper Allochthon. Marbles and schists showed large, flat-lying, isoclinal folds, refolded by broad, open, upright folds in a Caledonian transport direction. This is the end of the Ofoten synform, which closes against a major antiform running along the fjord. There is evidence of blocks sliding past each other at depth causing rotation of folds, as seen earlier in the Abisko area.

A few hundred metres away from the shore we found the base of the Øse Nappe again — so we are missing the Grønfjellet and Kvernmo nappes here. We saw very intensely strained phyllonites, in which biotite and chlorite predominate. We also found a boudinaged amphibolite dyke, with the phyllonitic fabric going around the boudins, showing that the amphibolite had already been stretched and sheared before the metamorphic event. The whole unit dips generally west. At this point Mark set us an exercise to determine the direction of movement. We dispersed along the outcrop and found several indicators. The most convincing for me are a series of folds verging top-to-the-east and a quartz calcite vein that shows signs of rotation in the same sense. This thrust has been dated at 383Ma, later than the main Scandian orogeny at 420–425Ma, and is evidence of post-orogenic collapse.

About 1km farther west we climbed down from the road to the shore where we found an exposure with large amphibole crystals in the form of *garbenscheifer* (sheaf slates) (Fig. 21, *overleaf*). These should not have formed, as they are out of equilibrium with the surrounding rock. They show both random orientation and alignment with the lineation (Fig. 22, *overleaf*).

Figure 21 Garbenschiefer form of amphibole porphyroblasts is reminiscent of the shape of a traditional sheaf of grain — south shore of Sagfjord (© Paul Speak 2011).

Figure 22 Kath and Lexy aligned parallel to lineation — Sagfjord (© Paul Speak 2011).

A short distance farther along we found garnet-mica schists of the Småtinden Nappe, the next unit above the Øse Nappe. The garnets are pre-tectonic, with no overprinting of strain, and have quartz envelopes relating to late stage shearing.

Our next stop was in the small village of Rotvika. As we got out of the cars a local lady greeted us and asked what we were doing, as "Nothing ever happens here!" The visit of a bunch of geologists was apparently the highlight of the week — or possibly the year! A major fault is visible at the western edge of the lake, the offset of a wide quartz vein showing it to be a normal fault. It is thought to be related to rifting at the break up of Pangaea. Jurassic and Cretaceous sediments have been deposited in the offshore basin but these have not yet been developed for oil. However, last year the Norwegian Petroleum Directorate paid all the fishermen in the area the equivalent of a year's income not to go to sea — while they conducted a major seismic survey, which would have disrupted fishing. Watch out for future developments.

The last stop of the day, 1km farther around the coast, was to see a series of metabasite dykes, which Mark said had a continental origin, unlike those seen on previous days, which had a MORB composition. With this we returned to our grass-topped hotel for dinner.

Figure 23 Artist Calle Örnemark's 30m high wooden sculpture Morgans Skip *at Lille Jamaica, Gratangen (© Linda M. H. Fowler 2011).*

Day 7 Lavangen to Ballangen (Lillevik en route) (1 August)

by Fran Kirby

We set off early to travel south to our next destination. It was a beautiful sunny day and we took a scenic route around Lavangenfjord and Gratangenfjord, stopping at several places to take photos. We spotted porpoises at the end of Lavangenfjord and there were spectacular views across the fjords towards Harstad.

Our first geology stop of the day was along the edge of Gratangenfjord, where we stopped next to what appeared to be the masts of a sailing ship, but on closer inspection turned out to be a sculpture created as a tribute to the 17th-century Welsh pirate and privateer Henry Morgan (Fig. 23). About 50yd (46m) from the sculpture there is an outcrop of marble that contains many grey to colourless, tabular crystals of zoisite. Zoisite is a calcium aluminium silicate that belongs to the epidote group of minerals and the crystals are formed when hot molten granite percolates through the marbles, causing alteration. The marble was in the upper level of the Gratangen Nappe and had undergone kyanite grade metamorphism.

The next stop on our way south was at a roadside exposure of gneisses. Here there is a large boudin structure, several metres across, of older gneisses, and within it there are traces of granite. The granite streaks within the boudin showed deformation that was possibly of Precambrian origin. Surrounding the boudin were younger, kyanite-bearing gneisses that showed a strong fabric, and had been deformed and metamorphosed during the Scandian Orogeny of 425Ma.

We travelled on into Narvik, the main town and port in the area, and after lunch went down to the port where we saw several large ships loading iron ore brought by rail from the Kiruna Mine in Sweden, which we had visited earlier in the week.

The main stop of the afternoon was near the harbour at Narvik, where we walked a short way along the beach to see part of the

Figure 24 Refolded fold with quartz saddle reef — Lillevik (© Linda M. H. Fowler 2011).

Lillevik Dyke Complex. This complex is thought to be the remnants of the sheeted dykes of an ophiolite, which have been highly sheared and deformed. At the first location there are bands of light and dark fine-grained mafic rocks that had been intensely folded and deformed. There had been multiple phases of folding with many complex folding structures, including isoclinal folds that had been refolded — folds within folds, lenses and saddle reef structures (Fig. 24). There are also crenulations and intense fabric parallel to the dykes and strong stretching lineations. There are fine-grained and coarser-grained metadolerites showing the dykes cutting through metasediments. Farther up the beach there are more mafic gabbros, also highly deformed and containing many complex folding structures. There are also large crystals of amphibole where the gabbro had been metamorphosed to greenschist facies amphibolite.

The Lillevik Dyke Complex is thought to be the remnants of an older ophiolite that had been subducted and then eroded and overlain unconformably by conglomerates. This resulting structure had then been deformed and metamorphosed during the Scandian Orogeny. The rocks have been dated to *c*. 470Ma and they have a MORB signature, indicating that the ophiolite may have formed as a spreading ridge in a back arc basin during the Caledonian. From here we continued to our destination and base for the next few days, a campsite in Ballangen.

Day 8 Ofotfjord and Hekkelstrand Quarry (2 August)

by Tamara Beck-Slinn

We spent this day at locations east of Ballangen in the Råna Massif, and then at a mine to the west.

Our first stop was at a road cutting in Råndal and Råna Mine, a few km east and up a side road into Råndal, a valley cut into the Råna Massif. This massif is a zoned pluton in an inverted cone shape (similar to Skågaard), intruded into high-grade gneiss and cutting its D1 deformation. Pressure and temperature studies show that this took place at 2kbar/6km; then both the gneiss and the intrusion were affected by deformation, which took place

Figure 25 Sheared conglomerate at Balsnes (© Linda M. H. Fowler 2011).

c. 12Ma later at 9.5 kbar/30 km. The road cutting has exposed several stages of cross-cutting norite dykes belonging to the Råna Massif, comprising augite norite cut by a dark blob of micronorite, and then cut by a dark titanium-rich norite and garnet-bearing trondhjemite. Dating of these indicates a major magmatic event prior to the Scandian (*c.* 425Ma), possibly related to the earlier Grenvillian back arc spreading. The likely origin of the dykes is an extensional phase affecting the crustal gneiss, which allowed decompression melting of the underlying mantle.

From here we retraced our route and, just before the campsite, turned inland up a track to Råna Mine. Originally mined for copper and nickel, the mine is currently used for aggregate, which is transported by conveyor belt and shipped from the nearby pier. There are 43.6Mt reserves of copper and nickel at 0.33% and 0.08%, respectively, but the present Danish owners (who also have a uranium mine in Spansdalen) hoped for platinum (which is present at < 30 parts per billion) and other rare earth metals, as China has now stopped their own export of these metals. Kuwait is in the last year of a seven-year lease for experimental drilling at the mine. We looked around and found a few pieces of bornite, but nothing more exciting. Ballangen has experienced environmental problems created by an iron mine farther south, which contaminated the fjord on a large scale; evidence can still be seen in the bright orange shore line. The bay has recently been infilled and grassed over.

From here we drove west through Ballangen and then turned down a farm track toward the coast at Balsnes, where the owner gave us permission to walk down to the foreshore. Here we saw an exposure at the base of the Evenes Group, which is thought to be an ophiolite fragment. A spectacular conglomerate (Fig. 25, *previous page*) similar to that already seen at Lavangen, containing very poorly sorted clasts of granite and quartzite, lies over a sheared mafic unit — the 'ophiolite' — which in turn is above the 150km long Øse Thrust.

Figure 26 Tremolite asbestos from Hekkelstrand quarry: this specimen is c. *120mm from top to bottom (© Linda M. H. Fowler 2011).*

Figure 27 Recent roof fall at Hekkelstrand (© Linda M. H. Fowler 2011).

Figure 28 Folded schist and dolomite at Hekkelstrand (© Linda M. H. Fowler 2011).

Our next location was Hekkelstrand marble quarry. Here we are in a minor plunging fold on the Ofoten synform and have moved up the stratigraphy into the Gratangen Nappe. A snow white, pure dolomite is quarried below ground by just three employees. Mark Anderson hopes a Plymouth University student will be taken on here to produce a three-dimensional model of the reserves for the next 10 years. The deposit is a bed in a major fold plunging to south and north in a saddle shape, with parasitic folds inter-bedded with schists and amphibolite. The main fold has a predictable shape but the smaller S-, Z- and M-folds, together with minor thrusts and faults, are not, and cause complications; hence the need for a three-dimensional model. In the board room we were shown samples of tremolite (one of six asbestos types), a fibrous alteration product (Fig. 26), and also of needle-like tremolite talc.

We walked into the mine and along one limb of the antiform where the dolomite has been extracted. The rock can be quarried out leaving little support: one 'room' was 70m × 40m, with a rock-bolted roof containing monitors to detect any movement. Although this is generally a safe mine, we were shown one small roof collapse that had happened a few days before (Fig. 27). At the midpoint of the antiform's 'saddle' a tunnel connects the two limbs and here, picked out by bands of schist, we were able to see the parasitic, isoclinal folds and faults (Fig. 28), together with

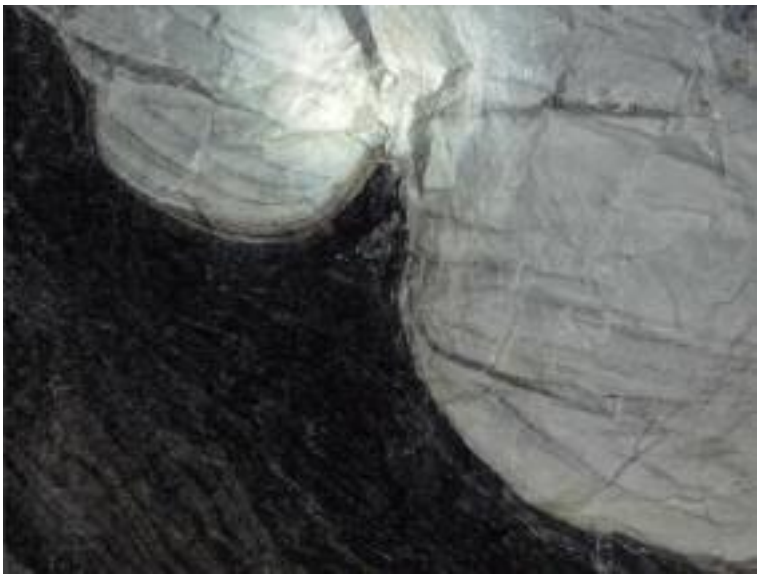


Figure 29 Mullions exposed in cross-section (© Linda M. H. Fowler 2011).

Figure 30 Amphibole porphyroblasts growing in random orientations near Bøstrand (© Linda M. H. Fowler 2011).

some spectacular mullion (flame) structures (Fig. 29) that complicate quarrying the deposit. The rock is quarried by drilling: for each blast 110.5m deep holes are drilled, three at a time using a 48mm drill and then blasted. The dolomite is used in cement, paint filler, pharmaceuticals and food additives.

Day 9 The Øse Thrust and Kjelde (3 August)

by Susan White

It was a damp, mild and misty day, and we were going to look for evidence of the path of the Øse Thrust after it descends the east side of the Ofoten syncline, to discover whether it continued up the other side. The Øse Nappe consists of inter-bedded marbles and schists. On the previous day, on the beach, we had seen the conglomerate outcrop at the base of the nappe. A mafic unit is below the conglomerate, and above the Øse Nappe is the Småtinden Nappe containing the Bø quartzite.

By 10.30 the weather was starting to clear, and we reached Bøstrand, which gives its name to the Bø quartzite and is on the eastern edge of the Ofoten synform. Our first stop was at a road cutting where the Tangen Thrust sheet was exposed to show metamorphosed turbidite fan sediments washed down from a carbonate shelf. The inter-bedded quartzites and garnet schists have isotope ages of *c.* 610Ma. Amphibolites had grown randomly across the grain, showing that they were formed after metamorphism (Fig. 30). The middle layer has thin quartz vein patches aligned with the lineation, and small carbonate pebbles with generally larger ones in the layer above. These pebbles had been flattened to ovals in the direction of the lineation (Fig. 31). These observations are consistent with deposition in a marine fan rather

Figure 31 Flattened carbonate clasts in metamorphosed turbidite fan sediments, Tangen Thrust sheet near Bøstrand (© Linda M. H. Fowler 2011).

Figure 32 The otters had scattered sea urchin shells such as this example across Punsvik foreshore (© Linda M. H. Fowler 2011).

than in a glacial till. If the pebbles had been deposited as glacial till, the clasts would be poorly sorted and angular. Its quartzite nature logically puts it with the formation of the Bø quartzite above it.

The second stop took us to the west side of the Ofoten synform, at the west headland of Punsvik Bay. The remains of shells, crabs and sea urchins left by otters littered the beach rocks (Fig. 32), and rock erratics had been littered by ice (Fig. 33). On the east side of the bay we saw an outcrop of the Hekkelstrand dolomite marble, mirroring the marble quarry exposure that we visited the previous day, which was on the west side of the Ofoten syncline, and shows that the marble folds right around the syncline.

From a position on the west headland of the bay the late Proterozoic Liland marble outcropped close by and can be seen continuing in the hills opposite, across Ofotfjord to the north. The bay rocks are the same garnet schists seen elsewhere in the Gratangen Nappe and had been extensively folded. They show a dip to the south-east into the synform, as expected. The granite veins toward the top of this nappe are characteristic of the beds above the Bø quartzite. These massive schists were formed from deep-sea clays, with some volcanics giving some more mafic areas. We saw a horizontal rainbow to the north, though none of us knew the explanation, other than someone's remembrance of more circular rainbows nearer the equator.*

The third stop was a bit farther west in an open area beside the road. The Bø quartzite here is banded, a result of its original bedding affected by tectonic movement. At an oblique angle a mineral lineation is developing in the direction of the travel of the thrust. The dip is 70°, once again to the south-east. Interbedded schists in the quartzite have garnets with strain shadows, further evidence of thrusting. The quartzite is cut by low-angle faults, perpendicular to the banding and showing the banding bowed around at the faults by the direction of shear. Farther back in the clearing there is a low cliff with the pattern of a duplex in the thrusting.

Between the harbour of Kjeldebotn and Kjelde, farther west along the coast, we passed several marbles of the Øse Nappe. The shoreline here has pink granite sand and grey marbles can be seen across the river mouth, again dipping south-east. The beach schists show highly deformed conglomerates that belonged to the Middle Allochthon, having a different form from the garnet-biotite schists seen previously. The conglomerate shows there is a stratigraphic gap between the top of the Middle Allochthon and the Bø quartzite seen at the previous stop, meaning that the Øse Thrust cut through those missing layers on the east side of the synform.

The fifth stop was at a quarry of Precambrian basement similar to that in the Rombak window in that it showed high-temperature strain features. The gneisses dip parallel to the Caledonian direction of transport, and had been aligned in that direction by the strain of shear contact. They dip south-east, into the synform, and there is a strong lineation at 110°, slightly oblique to the dip.

Figure 33 Ice-transported erratics at Punsvik (© Linda M. H. Fowler 2011).

** LF note: Wikipedia suggests that this phenomenon is seen when the sun is very high in the sky, never north of 55° (which is interesting considering we were at* c. *68° N), and cirrus clouds are present.*

Figure 34 Unloading and exfoliation of granite produces bare rock mountainsides near Bognes (© Barbara Norton 2011).

Plagioclase crystals in the gneiss show that the strain is top-to-the-right, that is, south-east. The gneiss is of granitic composition, and the quartz and biotite have recrystallised to smaller grains in response to the strain; and at similar temperatures to those in the Rombak window.

In conclusion, the basement was probably overthrust by the Middle Allochthon, and the Synform was formed by the middle to late Devonian brittle thrusting of the basement imbrications. The Øse Thrust is only found on the east side of the synform and continues at depth in the direction of the Scanobaltica subduction. The subduction of lower density crustal material results in its exhumation due to buoyancy. Isotope studies give an age of 425Ma for the stacking of allochthons on Baltica and 390Ma for the late stage thrusting of the basement.

Day 10 Ballangen to Sortland (4 August)

by Barbara Norton

Today we travelled from Ballangen campsite westward toward Vesterålen and the Lofoten Islands. The weather continued sunny, though cooler at 12°C. We left the allochthons and the Ofoten synform behind us and entered the Precambrian/Archaean basement. En route for the ferries, we saw pale grey granite slabs of bare rock, the result of weathering from exhumation exfoliation, where the reduction in pressure on uplift leads to large slabs of rock sliding off the mass (Fig. 34).

From the ferries, Skaberget to Bognes and then Bognes to Lødingen, we saw amazing views of the cloud-covered peaks.

From a lay-by at Lødingen we could see the Middle Allochthon to the east of the Tjeldsundetfjord, and the Archaean basement on the west, the contact running along the fjord. This contact is parallel to the Caledonian transport direction, and is similar to the major tectonic feature along Lake Torneträsk, which produced crossfolds in the allochthons. These Tysfjorden granites are dated 1.7Ga and are similar to those previously seen in the Rombak window. The Lofoten/Vesterålen Block is Precambrian and has been overprinted in the Cambrian: either it was originally part of Laurentia, in which case there must be a suture between it and the Baltic basement; or it is part of the Baltic basement, in which case there should be a nappe. The Lødingen granite shows deformation at 1.8Ga, during the Svecofennian event, which ties it to the Baltic basement.

Running almost north–south across Vesterålen is a magnetic discontinuity, an isograd. To the east are Archaean granites and low metamorphic grade gneiss amphibolites; to the west are the higher-grade granites and paragneisses of the Lofoten Islands, which also have a shallower moho. This discontinuity could be the Conrad discontinuity (where P waves show a jump in velocity) or have formed where the continental crust had been tilted. The simplest explanation is that the amphibolites originally overlaid the granites, which then rotated toward the east, leaving the amphibolites to the east and an almost vertical contact with the granites to the west. We travelled to Sortland through the amphibolites east of the discontinuity. At the end of Gullesfjord we could see the Gullesfjord granites, dated 2.6Ga, which have been intruded by the Lødingen granite, 1.8Ga. Farther on, the Gullesfjord granites have aligned mafic xenoliths showing crustal contamination, showing a shallow fabric with strong shearing. Again this fjord followed the contact between different granites.

Figure 35 The anorthosite-monzonite contact at Grøning. The upper, bare hillside is anorthosite, with a westward-dipping(left) contact clearly marked at the top of the more vegetated lower slopes on monzonite (© Barbara Norton 2011).

Still farther on we considered a more recent event, glaciation, discussing why the peaks were jagged. In the Dryas Stadial this area had a local ice sheet and valley glaciers, the peaks of the mountains remaining above the ice as nunataks. There has also been erosion related to uplift associated with rifting in the North Sea — the age of this uplift younging to the west. We then followed the north coast of Sigerfjord, where the highly sheared gneisses are steeply dipping.

Figure 36 A 200mm streak of pseudotachylite — south-east side of Eidsfjord (© Linda M. H. Fowler 2011).

After arriving at 'Sortland Camping og Motell' on Vesterålen we went west to Grøning on the Eidsfjord to see the high-grade metamorphism west of the isograd. As well as three otters, we saw a major contact of anorthosite over monzonite. This anorthosite had move westward and not in the Caledonian transport direction, in a normal fault (Fig. 35). From the Muscovites this movement has been aged at 403Ma. This movement shows there was extension in the Baltic basement at a time when it was also contracting farther east, shown by the Øse Thrust in the allochthons seen earlier.

On the way back we stopped at a road cutting to see pseudotachylite veins (Fig. 36). These long glassy streaks resulted from frictional melting during tectonic movement and have been dated as Devonian. These represent 'fossil' earthquakes that occurred deep in the basement, similar to deep earthquakes happening now in the Himalayas.

Day 11 Eidsfjord shear zone (5 August)

by Richard Buxton

We drove from Sortland 120 miles (193km) south-west, toward the southernmost part of the Lofoten Islands, to the Isle of Flakstadøy, through the beautiful scenery of peaks and fjords that had been formed in the Precambrian granulite zone of Lofoten. We saw sea eagles and some elk on the way.

On the beach at Myrland we observed monzonite (a plutonic rock with equal amounts of plagioclase and alkali feldspar and very little quartz), which graded into a sheared and altered zone, and then passed into amphibolite as we moved seaward and

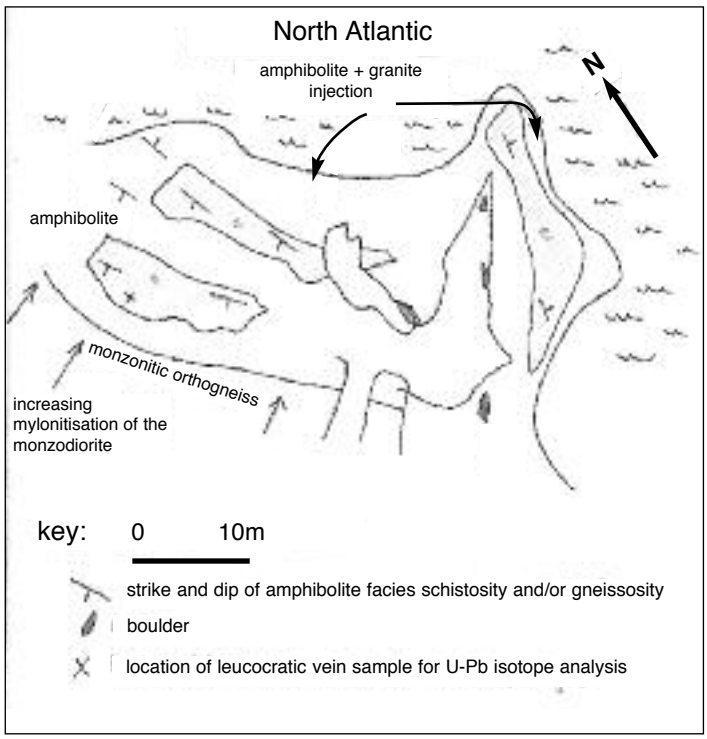


Figure 37 Field sketch of eclogite exposure on foreshore at Myrland (© Richard Buxton 2011).

to the north (Fig. 37). In the more mafic areas of amphibolite we observed small lenses of eclogite, consisting of green crystals of omphacite with some red garnets.

The interpretation of these rocks is that this part of Baltica had been subducted to deep levels, where eclogite had formed at pressures of 13Kbar (40–45km depth) and temperatures of 800°C to 900°C in shear zones that enabled fluids to promote the metamorphic reaction. This event must have occurred before the Scandian Orogeny (435Ma), because uranium dating in zircons

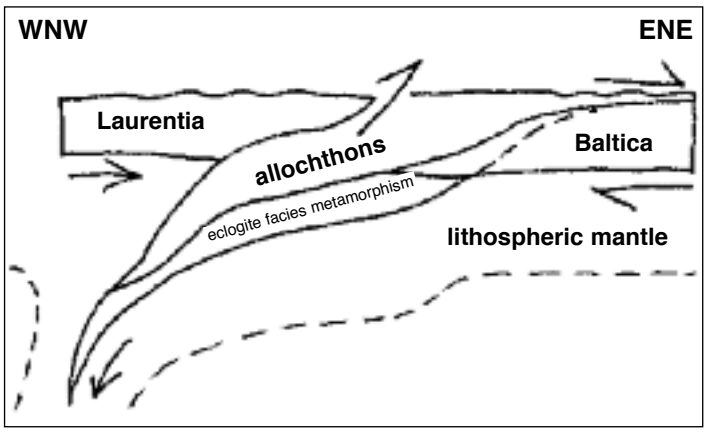


Figure 38 Formation of eclogite: later westward development of out-of-sequence thrusts provided a mechanism for exhumation (© Richard Buxton 2011).

showed two dates: 1.8Ga representing the protolith and 478Ma representing the re-setting when the eclogites formed (Fig. 38).

Later extrusion of this Baltic crust, from 405Ma to 370Ma, led to exhumation of the leading edge of Baltica, with eclogites in the Lofoten Block. Material was squeezed out like toothpaste as a thrust to the east occurred at the same time as extension to the west. This is a classic mountain-belt behaviour known as channel flow.

Day 12 Andenes and DIY (6 August)

by Linda Fowler

After the long drive of the previous day, this was a 'geology-free' day. The majority had planned a whale-watching boat trip from Andenes to the north, but sadly the weather was too rough. A smaller group of us drove around the island of Hadseløya, stopping at the Hurtigruten Museum in Stokmarknes and ate our lunch admiring some stunning views of the Lofoten Islands, where we had been on the previous day (Fig. 39).

Figure 39 The Lofoten archipelago from Utbjør on Hadseløya (© Linda M. H. Fowler 2011).

Figure 40 Gullesfjord granite quarry (© Gill Toney 2011).

Day 13 Sortland to Harstad (7 August)

by Gill Toney

On our final full day we moved from from Sortland to Bogen on the mainland. The first stop was at a quarry, half way up the western side of Gullesfjord, to examine the Gullesfjord granite, which, dated at 2.7Ga, is some of the oldest rock of the Lofoten Block (Fig. 40). There is evidence of S- and L-fabric, with strong linear fabric indicating strong stretching and prolate ellipses from flattening. This exposure is part of the Gullesfjord Shear Zone, an extensional structure in the Ofoten region, which has been dated, using recrystallised muscovite, to 370Ma (Mid Devonian). This is a major extensional episode that lasted some 30Ma, from 400Ma to 370Ma, and dips west, with a series of later brittle thrusts overprinting the original deformation, which dips north and is the equivalent of the Øse Thrust.

Taking the ferry from Flesnes to Refsnes we continued to travel north, but up the eastern side of Gullesfjord, then across to Harstad, the capital of Northern Norway. Harstad serves the Baltic and Norwegian seas and the oilfields. From a hilltop we looked northward beyond the town to Precambrian basement rocks, but a walk through the woods brought us to an exposed unit of a streaky, green-grey mafic rock with lineation running 110°–120° (WNW–ESE), and so is a Caledonian fabric that sits above the Øse Nappe. The unit is in fact the ophiolite that we first encountered near Narvik, and also at Lavangen and near Ballangen, although here it appears on the western side of the Ofoten synform. As at the other sites, an unconformity lies between the ophiolite and conglomerate units, but here the conglomerate is underneath rather than on top of the ophiolite, indicating that the units were overturned during the 430Ma Scandian event. The conglomerate contains extremely stretched granitic and mafic clasts. On the opposite side of the valley is a scarp slope dipping towards us with the two units the right way up.

In the afternoon we met up with Helen, one of Mark's students, at the Harstad campsite. After showing us her map of the area (Fig. 41, *opposite*), she took us to two of her mapping sites. Walking to the far side of a small bay we examined a unit containing intensely streaked out pebbles composed of different granite clasts, some feldspar rich, others quartz rich, with lineation to the south-east. This is the conglomerate unit above the mafic rock (ophiolite), so the units here are the right-way-up side of the fold. The shearing, however, is toward the north-west, which is not is expected for the Caledonian, but could imply that the basement is folded and shearing north-west. The outcrop is intensely folded, indicating high strain, and the mafic clasts are so strongly sheared that they form the matrix for the conglomerate. Thus the series of folds we observed both here and above Harstad stretch more than 2km and fold the conglomerate, mafic (ophiolite) and the basement, and should therefore be part of the Scandian deformation.

Figure 41 Helen's map of the Harstad area (© Gill Toney 2011).

In the next bay we examined a streaky and deformed 'ophiolite' with layers of white-grey and white-green gabbros with dark amphibolites between the layers, variations of fine to coarse grains and a strong stretching lineation. This ophiolite is part of the early Caledonian event, which, on erosion, formed clasts in the conglomerate and was deformed in the Scandian (Fig. 42, *overleaf*). Facing this unit is a green ultramafic rock with highly sheared round edges and angular folding verging east. This is also part of an ophiolite complex that has been metamorphosed to greenschist facies and serpentinite.

In the sequence of folds seen today, all of the large folds verge north-west while the local smaller folds verge to the east.

Our last stop was across the first bay from the conglomerate, to look at the unit above it, which also dips east-south-east. The rock looks like a granite and the foliation is deformed the same as the conglomerate, so how could clasts of this unit be incorporated into the conglomerate as it sits above it? There are numerous questions still to be answered. Is it an intrusion or is there a fault on the beach meaning the granite is older than the conglomerate? Or maybe the unit is not a granite, even though it contains quartz and feldspar. Could it be metasediments? or a granodiorite that has been thrust a distance and seems conformable? As the beach masks the contact what is this unit? In other places, this unit also dates to 470Ma. The conglomerate unit is thin in the bay and runs out at the top of the beach, so an irregularity is possible.

There is still much work to be done to unravel the full geological history of this fascinating area.

Conclusion

by Linda Fowler

After this fascinating climax to the trip we stopped overnight at Bogen, near the airport, and spent a leisurely morning before catching our flight down to Oslo and then onward to the UK. We had been tremendously fortunate not only with the splendid weather but also in our expert and knowledgeable leader who summarised the geology we had seen as follows:

Precambrian history is exposed at Kiruna and the Lofoten Islands, and is a marker for the front of the allochthons and Baltica. The youngest events consist of brittle faulting of the basement out to the coast in the Mesozoic and the development of the Norwegian Sea.

While the main phase of the Caledonian is known as the Scandian, which equates to the Grampian in the UK, the end of the Caledonian occurs post-Scandian and extends into the Devonian. The Baltic Shield is significantly shortened by thrusting. There are some structural highs (duplex structures) such as the Rombak window, which must be post-Scandian as

they fold the allochthons. At the same time as the shortening there is extension in the Lofoten Block because, as Laurentia and Baltica converge, Baltica should subduct but does not, leading to eastward thrusting and generating extension westward. The convergence also led to channel flow of weak, hot material that squeezed out laterally but did not deform the Precambrian basement crust as it is very rigid even though it was deformed itself.

During the Scandian event (430–420Ma) the units were heated in the subduction zone then thrust up and inverted, stacking higher-grade metamorphism on top of lower-grade. Most of the deformation occurred to the granite with kyanite and garnet grades forming.

The end of the early Caledonian event is marked by the unconformity between the mafic (ophiolite) and conglomerate units at *c.* 470Ma, with regional faulting and overprinting in the Scandian. The eclogite in the Lofoten Block (*c.* 470Ma) and eroded ophiolites indicate that Baltica was involved in an early collision event, but with a back arc or a microcontinent — a question still to be answered.

In the Grenvillian event, 1100Ma, only units on Laurentia were affected below the Øse Thrust, but these have been emplaced on top of the stack.

[LF note: There are more photos available on Flickr at http://flic.kr/s/aHsjvEpKPr a*nd a Google Earth kmz file of the localities is available for anyone interested.]*

Figure 42 Scandian deformation of the ophiolite at Harstad (© Gill Toney 2011).

OUGS Symposium 2011 field trip to Duke's Pass and Loch Katrine, Trossachs, central Scotland

John Mendum, BGS, Edinburgh

Purpose

This was a field trip on Saturday 20 August, as part of the OUGS Stirling Symposium 2011, to examine several exposures within the wide hinge zone of the Tay Nappe and to learn about some aspects of the overall three-dimensional geometry; to see small-scale examples of folds and related cleavages resulting from the three phases of deformation that have produced the regional-scale structure; to examine Dalradian metamorphosed sedimentary and volcaniclastic rocks of the Southern Highland Group, notably those of the Ben Ledi Grit, the Creag Innich Sandstone, the Loch Katrine Volcaniclastic and the Aberfoyle Slate formations.

It was led by Dr John Mendum (BGS Edinburgh).

Introduction

The Tay Nappe, which dominates the south-eastern part of the Grampian Highlands throughout their entire length, is the most extensive continuous structure in the Scottish Highlands. In fact it is one of the largest nappe structures in the world. Including its south-west extension into Northern Ireland, it is some 430km in strike length and up to 40km wide. Hence, it is not surprising that for many decades it has dominated discussion of the structural development of the Grampian Highlands.

Regional Structure

Figure 1a-b *(overleaf)* shows a simplified geological map of the Trossachs and surrounding areas of the Southern Highlands and Midland Valley and a generalised north-west–south-east cross-section from Crianlarich to Callander and Alloa. The cross-section shows how the overall Tay Nappe structure can be divided into two distinct parts: a 'Flat Belt', forming the sub-horizontal lower limb of the nappe, which dominates most of the Southern Highland Group outcrop in the Grampian Highlands; and a 'Steep Belt' (termed the Highland Border Steep Belt), which incorporates parts of the original hinge zone of the nappe and occupies a zone 5–12km wide, situated adjacent to the Highland Boundary. The monoclinal fold hinge that lies at the junction of the Flat Belt and Steep Belt is termed the Highland Border Downbend (HDB), commonly abbreviated to the 'Downbend'. This relatively late monoform has effectively folded down the 'nose' of the nappe to create the Steep Belt. As such it provides a cross-section across what were originally different structural levels of the Tay Nappe. Note that throughout almost the entire outcrop of the Tay Nappe, the present level of erosion either lies in the upside-down lower limb of the nappe or in its down-bent hinge-zone.

History of Research

The existence of a flat belt and steep belt has been known since the early days of geological mapping in the Highlands. However, with the exception of the Loch Tay Limestone, a stratigraphical succession within the nappe could not be defined or mapped and thus the geometry of the overall structure remained problematical. The first major breakthrough was in 1938, when S. M. K. Henderson identified the presence of a major anticline, which he named the Aberfoyle Anticline, within the stratigraphically enigmatic strata of the Steep Belt (Henderson 1938). Related structures were identified subsequently farther to the north-east by J. G. C. Anderson in 1947 (Anderson 1947). Note that both had been students at Glasgow University under E. B. Bailey, who had recognised nappe structures within the Dalradian succession of the Grampian Highlands some 20 years earlier. However, as the hinge zones of these regional-scale folds could not be recognised, it was unclear whether they were upward- or downward-facing (in modern terminology). Hence, the Aberfoyle Anticline was assumed to be a simple antiformal fold (i.e. upward-facing — the beds become younger in the structurally higher parts of the fold). A major breakthrough came in the 1950s, when R. M. Shackleton at Liverpool University, assisted by research students M. Stone and P. J. Stringer, developed the concept of 'facing', based on their work in the Aberfoyle area and wider studies in the Southern Highlands and Highland Border areas. In a seminal paper published in 1958, Shackleton showed how the relationships between bedding and cleavage can be used to determine the locations of antiforms and synforms in the absence of identifiable fold hinges, and that, when combined with 'way-up' determined from sedimentary criteria, these observations can define the facing direction of the major structures (Shackleton 1958). This breakthrough enabled the Aberfoyle Anticline to be recognised as a downward-facing synform. The basic model was applied to the whole hinge zone of the Tay Nappe, thus establishing the overall structural geometry of the Southern Highlands to be much as we know it today.

Development of the Tay Nappe — questions

But that was only the first step. How and when did such an enormous structure originate and develop? What does it tell us about the overall structural evolution of the Caledonian mountain belt? These are the questions that have exercised the minds of numerous geologists over the past 50 years and there is still much controversy today. Some key papers have been Harris *et al.* (1976), Bradbury *et al.* (1979), Krabbendam *et al.* (1997), Mendum and Thomas (1997) and Rose and Harris (2000). In particular, what was the nature of the right way-up upper limb of the nappe before it was removed by erosion? Did it even exist at all? Some of these questions were touched upon during the afternoon's excursion.

Geology of the Trossachs area

Structure and sequence

In the Loch Lomond–Callander area the hinge zone of the Tay Nappe has been divided into three main folds: the Ben Vane Synform, the Ben Ledi Antiform and the Aberfoyle Anticline (synform). These downward-facing structures occupy the Steep Belt and are truncated at their south-eastern margin by the Highland Boundary Fault (HBF), along which lie scattered fault-bounded outcrops of Arenig age black shaly meta-mudstones, cherts, limestones, conglomerates, turbiditic arenites and mafic lavas, juxtaposed with several serpentinised ultramafic bodies. These lithologies, collectively termed the Highland Border Complex, form narrow lenticular outcrops, commonly widely

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Figure 1 Geological map of the Trossachs and surrounding areas (from Browne, M. A. E. and Mendum, J. R. SNH/BGS).

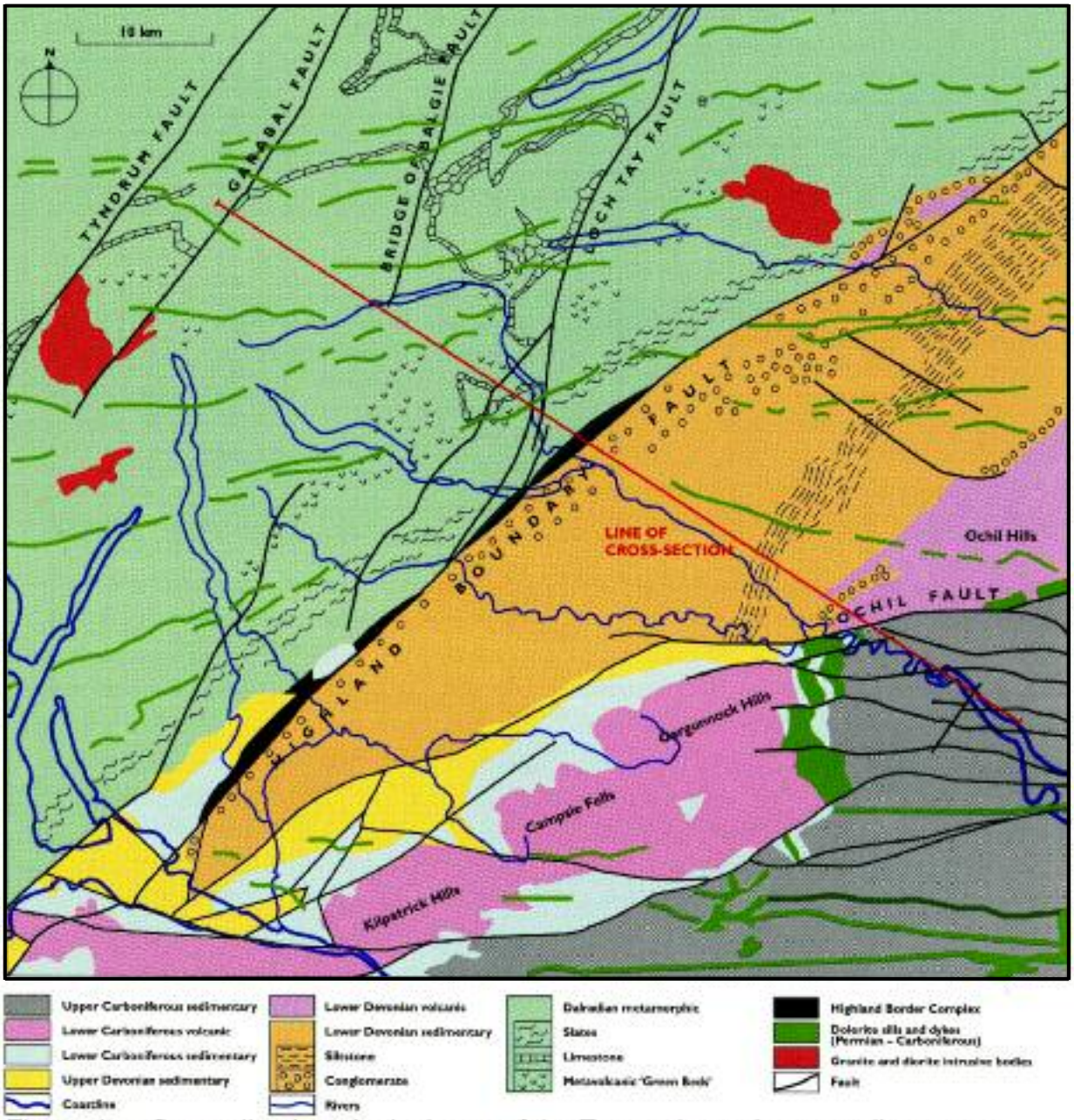


Figure 1a - Generalized geological map of the Trossachs and surrounding areas (From Browne, M.A.E. and Mendum, J.R., SNH/BGS).

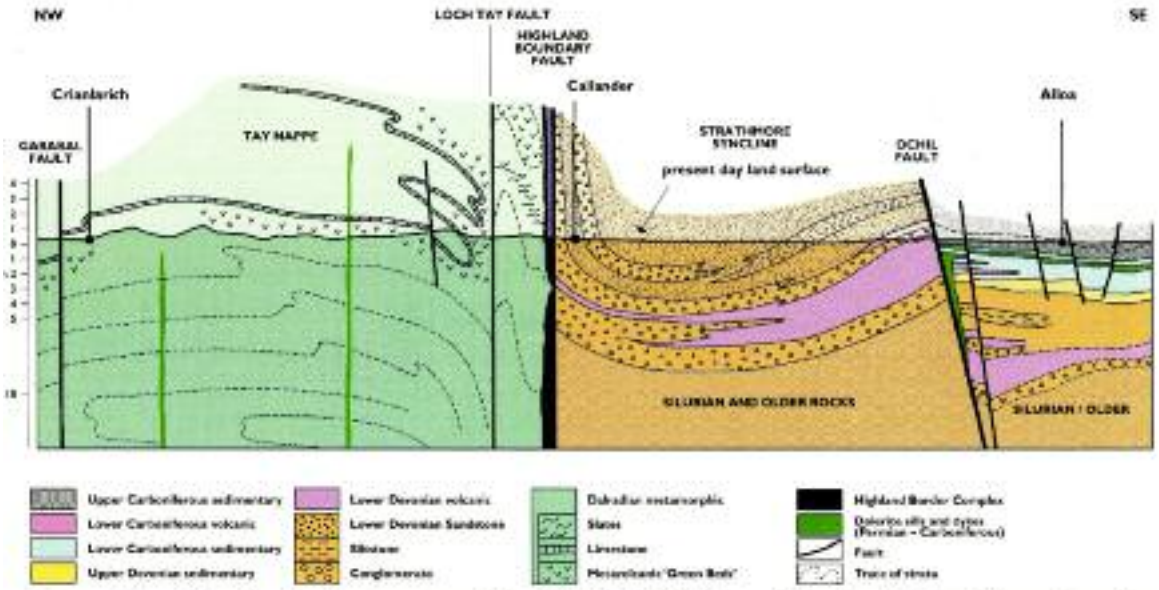


Figure 1b - Geological cross-section (NW - SE) from Crianlarich to Alloa showing the overall structure of the Southern Highland and Midland Valley. (From Browne, M.A.E. and Mendum, J.R., SNH/BGS).

separated, that appear to have been 'smeared' along the Highland Boundary in both Scotland and Ireland. Note that although the HBF now forms the structural boundary to the Tay Nappe, the fault must conceal an originally more complex continental margin, suture, and/or subduction zone that existed in Cambrian–Ordovician time during the formation of the Tay Nappe.

The Dalradian rocks of the Trossachs area comprise interbedded gritty arenites (meta-sandstones), meta-siltstones (semipelites), meta-mudstones (pelites) and volcaniclastic sandy and silty units ('Green Beds'), all assigned to the Southern Highland Group. The stratigraphical succession is varied even within the area of the 1:50 000 sheet 38E (Aberfoyle). Figure 2 *(overleaf)* shows part of this 1:50 000 sheet; note that the keys to the Dalradian lithostratigraphy are shown in Figure 3 *(page 91)*. An almost north–south trending cross-section across this area showing the generalised structure and lithostratigraphy is reproduced in Figure 4 *(page 92)*. These figures give some idea of the real complexity found here, partly a product of the Grampian ductile deformation, but also as a result of the Trossachs–Duke's Pass area being traversed by the Loch Tay Fault and its related splays. The faults, which mostly post-date formation of the Tay Nappe and of the Downbend, show both lateral and vertical movements and have effectively segmented the geology, making it difficult to trace both individual stratigraphical units and the traces of the major and minor fold structures.

Lithostratigraphy

The Loch Tay Limestone, the youngest formation in the Argyll Group, mainly represents shallow shelf accumulation of carbonate sediment just prior to rift-drift and the formation of the Iapetus Ocean (*c.* 580–515Ma). In contrast, the overlying Southern Highland Group rocks were deposited in deeper marine conditions, mainly by turbidity currents on the subsiding continental margin and continental slope. In the early stages of deposition significant facies variations developed, reflecting the rapid subsidence of parts of the continental edge, formation of fault scarps, and growth of turbiditic fans. Coarse-grained, locally gritty, and even conglomeratic arenites were deposited rapidly in channels on the continental slope and in the more proximal inner zones of deep-water submarine fans, whereas the finer grained pelitic sediments were mainly laid down as overbank deposits or in the more distal parts of the fan. Hence, lateral continuity of individual lithological units is rare, particularly in the older parts of this succession. Sedimentation was accompanied by mafic volcanism that resulted in intrusion of sheets and lenticular bodies of dolerite, and the input of volcanic detritus into the sand-silt-mud-dominated turbiditic succession.

The higher units of the Southern Highland Group, notably the Ben Ledi Grit Formation, show less facies variation and are generally more pelitic. They represent longer time periods with possible gaps in the sequence. The Leny Limestone and Slate Member of the Keltie Water Grit Formation that lies near the top of the succession contains deep water trilobites that date it as uppermost Lower Cambrian (*c.* 515Ma).

The Southern Highland Group

The oldest unit of the Southern Highland Group represented in the Trossachs area is the Aberfoyle Slate Formation, a series of dark-grey to green-grey and maroon slaty pelites and semipelites, formerly quarried extensively on the slopes of Creag Innich, *c.* 1km west of Duke's Pass. The formation occupies the hinge zone of the Aberfoyle Anticline (*see* Fig. 4) but cannot be recognised farther north around Loch Katrine, where the Loch Katrine Volcaniclastic Formation occurs at a similar stratigraphical level. Arenite-rich units, such as the Ardnandave Sandstone Formation and Creag Innich Sandstone Formation, have been delineated in these lower parts of the Southern Highland Group and can be mapped over restricted areas. As such they appear to represent relatively limited high-energy sandy influxes into the turbiditic palaeoenvironment.

The overlying Ben Ledi Grit Formation is thicker and more extensively developed. It consists of interbedded arenites, wackes, semipelites and pelites and locally shows features that can be matched in the classic Bouma sequences typical of turbiditic successions (grading, pebbly bases, cross-bedding, etc). Its more lithologically variable bedded nature has resulted in complex patterns of cleavage development and small-scale folding.

West of the Loch Tay–Duke's Pass Fault system the succeeding Loch Ard Grit Formation is dominated by arenites with slaty pelitic interbeds, and is in turn overlain by the Bofrishlie Slate Formation. This uppermost unit consists of dark grey to black, slaty graphitic pelite and semipelite, locally with minor thin sandstone beds. In contrast, east of the fault system the Keltie Water Grit Formation overlies the Ben Ledi Grit Formation. It consists of pale grey to white more quartz-rich arenites, locally gritty, with interbedded thin pelitic units. The unit still shows evidence of its has turbiditic origins but in its upper part it includes the Leny Limestone and Slate Member, composed of graphitic pelites and semipelites with thin dark grey meta-limestones containing trilobites.

Structural history

The structural history of the Trossachs area can be considered in terms of three of the four principal phases of deformation that can be recognised regionally across the Grampian Highlands. These are termed D_1, D_2 and D_4. Note that D_3 structures are well developed elsewhere in parts of the Tay Nappe, but are absent here.

The first two of these, D_1 and D_2, were associated with prograde metamorphism during the generation of the Tay Nappe, whereas D_4 is associated with retrograde metamorphism during or following later uplift of the Tay Nappe and the formation of the Downbend (HBD). D_1 resulted in the formation of large-scale folds of bedding (F_1) associated with a fanned spaced cleavage (S_1) in meta-sandstones and a slaty cleavage in the finer grained more pelitic beds. At high levels of the nappe exposed in the Steep Belt, these folds are preserved as downward-facing structures with steep north-west-dipping to sub-vertical axial planes. As noted above, in this area there are three separate kilometre-scale F_1 folds developed: the Ben Vane Synform, the Ben Ledi Antiform and the Aberfoyle Anticline (synform) (Shackleton 1958; Mendum and Fettes 1985). As the HDB lies close to the hinge of the structurally lowest fold, the Ben Vane Synform, the Steep Belt here is occupied by the complex hinge of the Tay Nappe (*see* Figs 1 and 4). Hence, the entire succession effectively youngs downwards and to the south-east in the Steep Belt.

The second deformation (D_2) was only developed at intermediate and low structural levels of the nappe. It was coincident with the peak of metamorphism and most authors have interpreted it as having a strong simple-shear component. What is in dispute is the role of the D_2 deformation in modifying or generating

Figure 2 Abstract from the 1:50 000 bedrock map for Sheet 38E (Aberfoyle) (BGS 2004), showing the Duke's Pass and Loch Achray areas, and the eastern part of Loch Katrine. For keys to the Dalradian succession see *Fig. 3.*

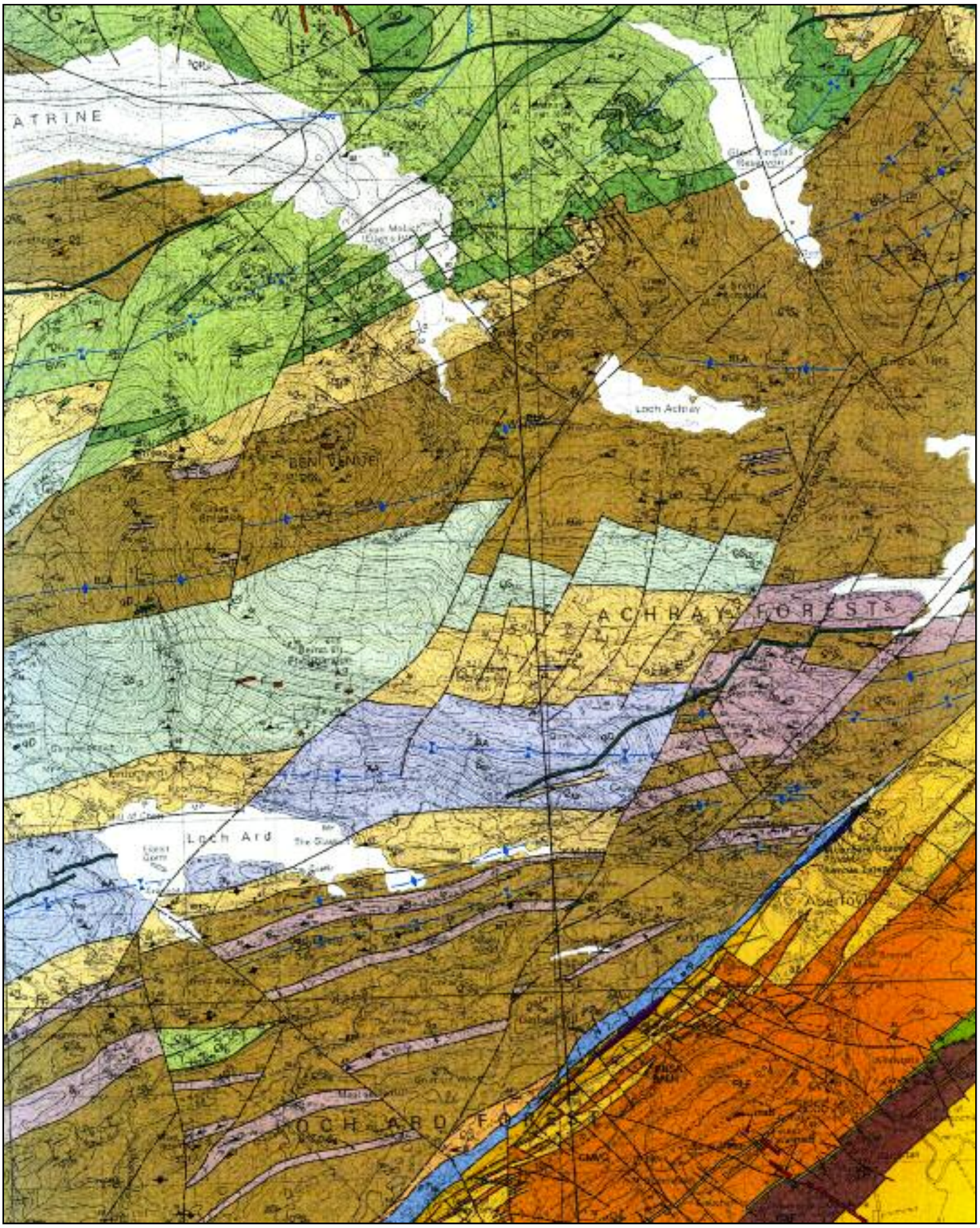
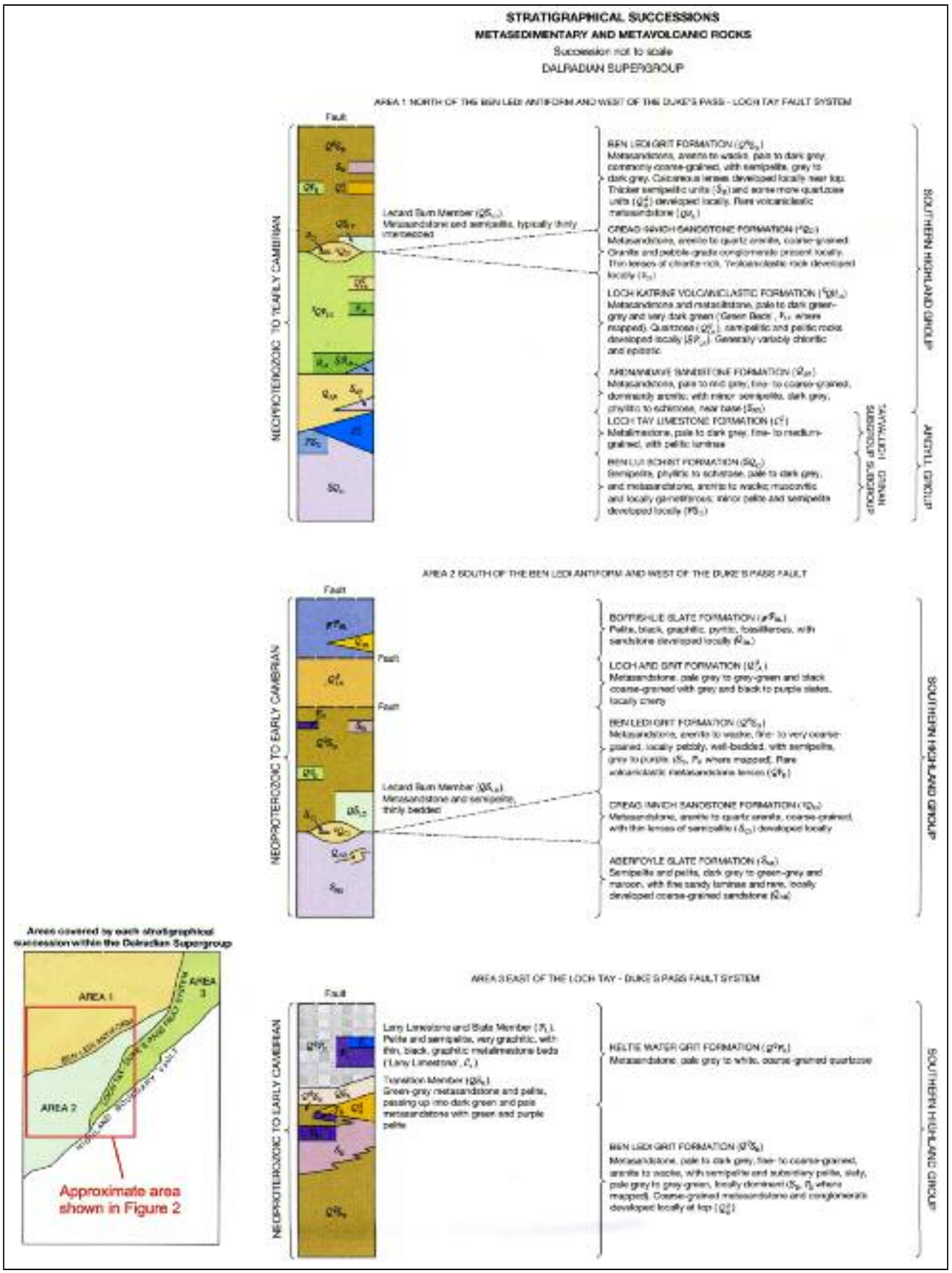


Figure 3 Keys to the bedrock lithostratigraphy of the Dalradian Argyll and Southern Highland Group rocks on Sheet 38E (Aberfoyle) (BGS 2004).


STRATIGRAPHICAL SUCCESSIONS
METASEDIMENTARY AND METAVOLCANIC ROCKS
Succession not to scale
DALRADIAN SUPERGROUP
AREA 1 NORTH OF THE BEN LEDI ANTIFORM AND WEST OF THE DUKE'S PASS - LOCH TAY FAULT SYSTEM
Fault
NEOPROTEROZOIC TO EARLY CAMBRIAN
Ledard Burn Member
Metasandstone and semipelite, typically thinly interbedded
BEN LEDI GRIT FORMATION
CREAG INNEICH SANDSTONE FORMATION
LOCH KATRINE VOLCANICLASTIC FORMATION
ARDNANDAVE SANDSTONE FORMATION
LOCH TAY LIMESTONE FORMATION
BEN LUI SCHIST FORMATION
SOUTHERN HIGHLAND GROUP
TAYVALLICH SUBGROUP
ARGYLL GROUP
AREA 2 SOUTH OF THE BEN LEDI ANTIFORM AND WEST OF THE DUKE'S PASS FAULT
Fault
Fault
Fault
NEOPROTEROZOIC TO EARLY CAMBRIAN
Ledard Burn Member
Metasandstone and semipelite, thinly bedded
BORRISHLIE SLATE FORMATION
LOCH AND GRIT FORMATION
BEN LEDI GRIT FORMATION
CREAG INNEICH SANDSTONE FORMATION
ABERFOYLE SLATE FORMATION
SOUTHERN HIGHLAND GROUP
Areas covered by each stratigraphical succession within the Dalradian Supergroup
AREA 1
AREA 2
AREA 3
Approximate area shown in Figure 2
AREA 3 EAST OF THE LOCH TAY - DUKE'S PASS FAULT SYSTEM
Fault
NEOPROTEROZOIC TO EARLY CAMBRIAN
Lany Limestone and Slate Member
Pelite and semipelite, very graphitic, with thin, black, graphitic metalimestone beds (Lany Limestone,)
Transition Member
Green-grey metasandstone and pelite, passing up into dark green and pale metasandstone with green and purple pelite
KELTIE WATER GRIT FORMATION
Metasandstone, pale grey to white, coarse-grained quartzose
BEN LEDI GRIT FORMATION
SOUTHERN HIGHLAND GROUP

the Tay Nappe and the overall direction of the sense of shear. Large-scale F_2 folds are only rarely found in the Southern Highland Group, but small-scale folds, generally with a north-westerly vergence, are abundant. These F_2 minor folds, notably of the S_1 spaced cleavage striping, typically show thickened hinge zones and moderately to highly attenuated limbs. Associated with the D_2 deformation is the formation of a second spaced cleavage (S_2), which typically is orientated near-parallel to bedding. This S_2 cleavage is dominant in the Flat Belt of the Tay Nappe.

The D_4 deformation occurred considerably later, either accompanying the uplift that resulted in the formation of the Downbend, or more probably, subsequent to Downbend formation. The characteristic minor F_4 open folds and associated north-west-dipping S_4 crenulation cleavage are absent close to the Highland Boundary but become well developed as the Downbend hinge is approached. These D_4 structures continue across the Flat Belt and are dominant in the Ben Lawers area, where they relate to the large-scale Ben Lawers Synform.

The field excursion

We started at the David Marshall Lodge Visitor Centre by Aberfoyle where the regional geological picture and the overall geometry of the Tay Nappe were explained together with some of the important structural principles involved. The three localities visited on the excursion had been chosen to demonstrate the increasing structural complexity and metamorphic grade in the hinge zone of the Tay Nappe, compatible with their greater tectonic depth prior to the formation of the Downbend (HBD) (*see* Figs 2 and 4).

We started at a high structural level at the Duke's Pass, where the early D_1 deformation and folding were seen, largely unsullied by later deformation phases, and then moved on to view lower structural levels in the nappe exposed at Craig Noran, and subsequently by the Trossachs Pier at Loch Katrine. At these two latter localities, the overprinting by the D_2 fabrics and related rotational and strain effects were observed.

The first two localities form part of the Duke's Pass GCR site, and the Loch Katrine localities are on the public road section and by the car park. Thus, we were requested to use our geological hammers only for pointing and as scales for photography. Also we were reminded to take care when traversing along or across the narrow Duke's Pass road (A 821).

Hill Cottage [NN 5158 0294]

Firstly, it was pointed out that the trace of the north-north-east-trending Duke's Pass Fault is mapped as passing through the cottage and down the burn to the south-south-west. The fault has downthrown the beds to the east-south-east (*see* Fig. 2). To the west-north-west of the fault lies the Aberfoyle Anticline, a relatively simple, kilometre-scale synformal structure with the Aberfoyle Slate Formation occupying its hinge zone. In contrast, to the east-south-east of the fault the fold structure is more complex with several medium-scale antiforms and synforms developed in the arenites and slaty pelitic rocks of the Ben Ledi Grit Formation.

We visited some vegetated outcrops 150–200m south-east of Hill Cottage, just on the north-east side of the A821, adjacent to a small glacial meltwater channel. Here, we noted that the arenites were gritty with coarse-grained pebbly bases and grading

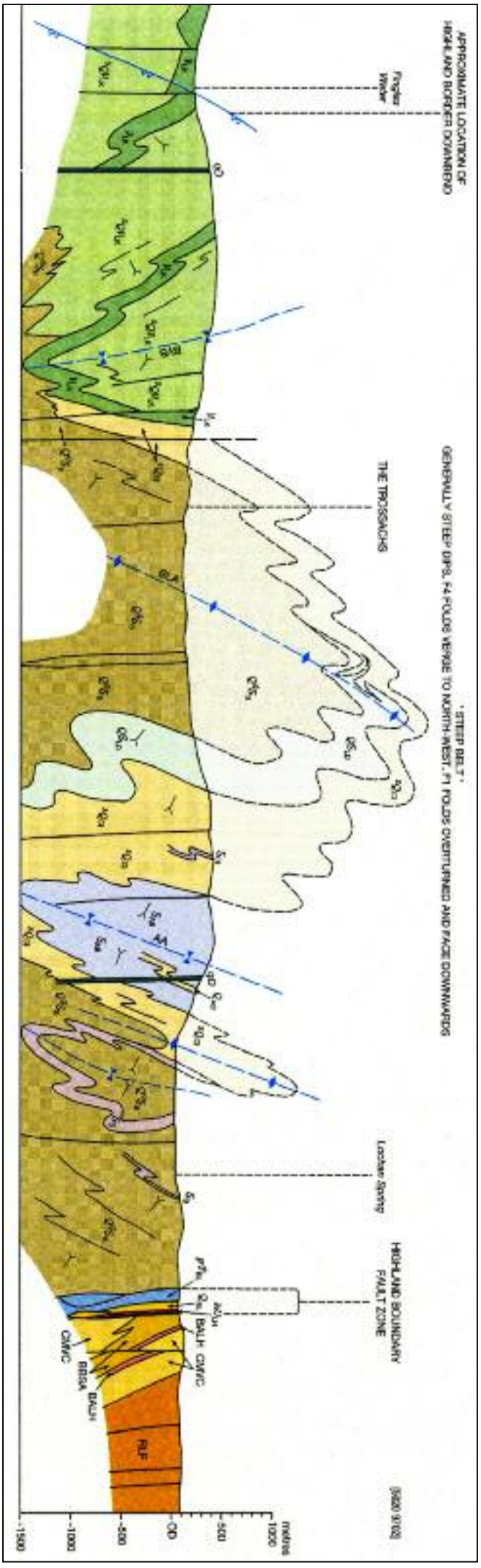


Figure 4 Cross-section (trending 358°–178°; see Fig. 2) across the Loch Katrine–Loch Ard forest district, showing the structure of the hinge zone of the Tay Nappe within the Highland Border Steep Belt (abstracted from 1:50 000 bedrock map for Sheet 38E (Aberfoyle) (BGS 2004).

present locally, but in some outcrops bedding and structural relationships were difficult to discern. The bedding mostly dips moderately to steeply to the north and north-east, with the grading and gritty bases showing that the beds are certainly inverted in places. Although several antiforms and synforms are present in these outcrops clear hinge zones are not readily seen and it is only the relationships between bedding and the S_1 spaced and slaty cleavage that indicate the actual geometry. We did satisfy ourselves that the beds faced downwards to the southeast and that the structures were almost entirely the result of D_1 deformation. Also that metamorphic grade was low, compatible with upper greenschist facies conditions.

Craig Noran [NN 5046 0662]

From Hill Cottage the A821 at first runs north-eastward along the trace of the Duke's Pass Fault, but then turns west and north-west and crosses the structural strike. Hence, at Craig Noran we were able to observe deeper structural levels, but still staying within the outcrop of the Ben Ledi Grit Formation. The cliffs expose clearly bedded arenites and interbedded pelitic units that dip 10° to 50° to the north and north-west. Some indications of grading were seen in parts implying that the beds were inverted. The outcrop lies just north of the trace of the hinge of the Ben Ledi Antiform (*see* Figs 2 and 4). Note that in the Glasgow Excursion Guide (Bluck and Lawson 1992) the description states that Craig Noran lies within the Flat Belt, but this is incorrect. The S_1 spaced pressure solution cleavage is prominent here, but is typically folded by a range of pervasive, small- to medium-scale, open to tight F_2 fold structures. Pressure solution cleavage developed, notably in the arenites, during both D_1 and D_2 deformation accompanied by metamorphism under greenschist facies conditions. The cleavage is generated when new micas and chlorite grow in defined narrow cleavage domains, typically initially centimetre-spaced, in concert with local migration of quartz into the intervening 'microlithons'. When traced laterally these microlithons coalesce and ramify in detail, such that the rock between the cleavage domains forms a series of elongate quartz-rich three-dimensional lenticles. When spaced cleavage has been subjected to differential weathering, the lenticular and more silica-rich microlithons tend to stand proud, with the cleavage domains forming narrow linear indentations. Spacing of the pressure solution cleavage depends on metamorphic conditions, the availability of fluids, and deformation prevalent at their time of formation. In the Dalradian Southern Highland Group rocks the S_1 cleavage has commonly become attenuated, reflecting the locally extended D_1 deformation history. However, the subsequent D_2 deformation has also been variably superimposed on the D_1 structures giving rise to both further attenuation of S_1 and small-scale F_2 folding of S_1 dependent on the orientation of the D_1 structures. Hence, it is difficult to interpret the relative roles of the D_1 and D_2 events in most outcrops. The axial planes of the F_2 folds generally lie nearly parallel to the bedding in the Southern Highlands. At Craig Noran only rarely is a new S_2 spaced cleavage developed.

The Trossachs Pier car park [NN 4957 0710]

At the south end of the Trossachs Pier car park, a short distance above the reed bed the wooded crags show complex bedding, cleavage (S_1 and S_2) and minor F_2 fold features (Fig. 5). They provide an example of the structurally complexity generated when both D_1 and D_2 deformation features are developed in mixed arenite-pelite lithologies, here again belonging to the Ben Ledi Grit Formation, but in its stratigraphically lowermost part. Similar structures are found across much of the Southern Highlands from Stonehaven to Kintyre. The bedding here dips moderately steeply north-west, and although it is inverted, convincing examples of grading are difficult to find. However, the nature of both S_1 and S_2, and variations in their intensity, does reflect the original turbiditic lithologies, showing up the coarser, more quartz-rich arenite beds, and those with a greater silt and mud content. The S_1 cleavage dips more steeply to the north-west than bedding but it has been deformed by D_2. Locally, it is folded by open to tight north-west-verging minor F_2 folds with the associated development of a secondary S_2 spaced cleavage, whose spacing ranges from *c.* 5mm to several centimetres We noted that the axial planes to the F_2 folds and the related S_2 cleavage were orientated sub-parallel to bedding. D_2 effects are more dominant at this locality — a result of passing down to yet deeper structural levels. These outcrops lie on the common limb between the hinge zones of Ben Ledi Antiform and Ben Vane Synform. Note that if the effects of the Downbend are reversed the beds here would be right-way-up (*see* Fig. 4).

Figure 5 Steep north-westerly dipping bedding (by hammer) in arenites and subsidiary pelitic units of the Ben Ledi Grit Formation showing S_1 and S_2 cleavages and minor scale F_2 folding of S_1 — crag overlooking the south-west end of the car park by the Trossachs Pier.

Loch Katrine [NN 4954 0724 to NN 4922 0781]

The final localities were reached by a walk along the road on the Loch Katrine side from the Pier to the water outlet into Loch Katrine from the Glen Finglas Reservoir. We started in the lowermost part of the Ben Ledi Grit Formation and our horizontal traverse took us stratigraphically downwards through the more arenite-rich Creag Innich Sandstone Formation, with its thicker beds and gritty to conglomeratic horizons, and into the Loch Katrine Volcaniclastic Formation. Although there are numerous road cuts along the loch side, their condition is variable — some are geologically informative, others much less so. Structurally we are again to the south-east of the hinge zone of the Ben Vane Synform, but on this section there is evidence of large-scale F_1 and possibly F_2 folding, probably reflecting the both thicker nature of the beds and the increased competence of these formations. As above, the beds are generally inverted, but if corrected for the Downbend they would form part of a right-way-up fold limb facing south-east within the overall hinge zone of the Tay Nappe.

The first outcrop just past the kiosk shows steeply north-north-west-dipping bedding, S_1 and F_2 / S_2 relationships in arenites of the Ben Ledi Grit Formation. Some of the beds may have a minor volcaniclastic component. Although apparently complex, once the individual elements (particularly the bedding) have been recognised, its structure can be more readily assessed. Farther on, bedding and S_1 and S_2 can be recognised in the locally gritty, thickly bedded arenites of the Creag Innich Sandstone Formation and there are good examples of S_1 folded by F_2. Around [NN 4950 0765] clear bedding, dipping gently to the north-north-east is defined by in coarse-grained arenites and interbedded thin pelites with S_1 orientated at a high angle to bedding. Clearly we are in an F_1 fold hinge here, albeit with some superimposed D_2 deformation and minor F_2 folding and cleavage development (Fig. 6).

Figure 6 Bedding–cleavage relationships in arenites and semipelites of the Creag Innich Sandstone Formation — Loch Katrine side.

Farther west there are more massive arenites in which both bedding and cleavage traces are difficult to discern. Also in some more mixed arenite-pelitic units possible examples of open to close F_4 folding can be seen. These late-stage structures fold the now dominant S_1–S_2 cleavage, bedding and quartz veins and generally have north-west dipping axial planes, with local development of a crenulation or fracture cleavage. The related metamorphic assemblages are only at mid greenschist facies. F_4 folds and S_4 cleavages become more common as the Downbend hinge is approached.

From the loch side road we turned east to walk a short distance up by the Glen Finlas Reservoir outlet, where a formerly quarried face exposes typical volcaniclastic arenites ('Green Beds') of the Loch Katrine Volcaniclastic Formation. We appreciated their massive nature and distinctive but subtle bluish green-grey colour imparted by their significant chlorite and epidote content. Bedding could be discerned but evidence of the S_1 and S_2 cleavages was characteristically absent.

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The zeolite minerals (part 1)

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Introduction

Axel Fredrick Cronstedt, born in 1722, was a Swedish chemist and mineralogist who described a new mineral from the Svappavarra Copper Mine in northern Sweden. The new discovery was subsequently named as 'stilbite' and recognised by its 'intumesence' — the visible loss of water when heated (Cronstedt 1756). He coined the word 'zeolite' to name this new mineral from two Greek words, ζείν and λιθοσ, meaning 'boiling stones'. It proved to be a member of a class of hydrated aluminosilicates. [Note: the material from Svappavarra has been re-investigated and shown to be stellerite — a calcium-rich form of stilbite (Collela and Gualtieri 2007).]

Cronstedt later made a fundamental contribution to mineralogy in that he was the first person to point out the difference between minerals and rocks as composites of various minerals, a turning point in the history of mineralogy. He was the first man to discover the element nickel, as well being responsible for the introduction of potatoes and other crops into his native region of Dalarne, Sweden (Schlenker and Kühl 1993).

Zeolites comprise building blocks composed of four tetrahedrally arranged oxygens with silicon or aluminium at their centres. The tetrahedra are linked to each other by all four of their corners to create a continuous framework, as illustrated in Figure 1. The difference in cation charge between Si^{4+} and Al^{3+} leaves a residual negative charge on the oxygen framework, which is compensated by the presence of cations (commonly Na, K, Mg, Ca, Ba) within the open cavities and channels in the framework. In some zeolites the volume available to these cations is as high as 50% of the total crystal volume. This means that the zeolite framework has also plenty of vacant volume to accommodate water molecules, whose easy loss causes the intumesence.

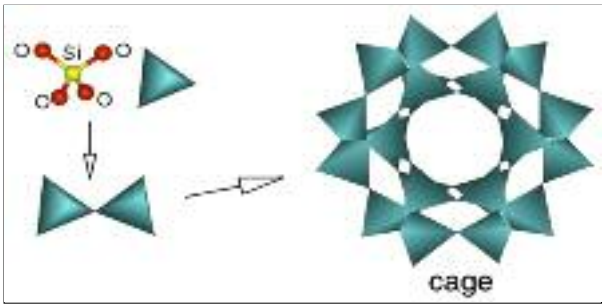

Figure 1 The assemblage of silicon [Si O4]$^{4-}$ and aluminium [AlO4]$^{5-}$ tetrahedra to create continuous cage structures.

In most zeolites the cations are highly mobile and can be readily replaced by other cations in solutions in contact with the zeolite. This is the property of cation exchange. Early publications often credit zeolites as being responsible for the uptake of the ammonium cation by soils and also as the agency in early domestic water softeners. Both these points are incorrect. Clay minerals are the source of ammonium uptake by soil, and the first water softeners were synthetic amorphous aluminosilicates. All zeolites are highly crystalline solids and they are much collected for their beauty (Figs 2–7, *overleaf*) and appear in many museum collections.

In the modern era someone with a desire to learn more about zeolites might be drawn to a web search. This would yield millions of hits, as publications about zeolites run to about a thousand every month! This is because Professor Richard Barrer, working in the UK between 1934 and 1993, recognised the industrial potential of zeolites and developed laboratory-scale syntheses, which formed the basis of the plant scale manufacture developed by scientists from the Union Carbide Corporation in the 1950s. The industrial use of these synthetic zeolites and their natural cousins is described later.

Estimation of the numbers of known zeolites is rendered difficult by the continuous discovery of new species — both natural and synthetic. Suffice it to say that it has been calculated that there are thousands of spatial arrangements of linked tetrahedra that can give rise to zeolite frameworks. A further complication has come from the discovery of zeolitic frameworks, both in nature and in the laboratory, that have elements other than silicon or aluminium in the tetrahedral positions in their frameworks (*see* Fig. 9). These positions can be occupied by boron, beryllium, phosphorus, cobalt, iron and many other elements. One unique aspect about the zeolite minerals is that some species first prepared and characterised in the laboratory have been subsequently found in nature, while some other well-known synthetic

Table 1 Natural zeolites

Amicite	Gismondine*	Natrolite*
Ammonioleucite	Gobbinsite*	Offretite*
Analcime*	Gonnardite*	Pahasapaite[d]
Barrerite	Goosecreekite	Partheite
Bellbergite	Gottardiite	Paulingite*
Bikitaite	Harmotome*	Perlialite
Boggsite	Heulandite*	Phillipsite*
Brewsterite*	Hsianghualite[a]	Pollucite
Chabazite*	Kalborsite	Roggianite[e]
Chiavennite[a]	Laumontite*	Scolecite*
Clinoptilolite*	Leucite	Stellerite*
Cowlesite[b]	Levyne*	Stilbite*
Dachiardite	Lithosite	Terranovaite
Direnzoite	Lovdarite[a]	Thomsonite*
Edingtonite*	Maricopaite	Tschernichite
Epistilbite*	Mazzite	Tschörtnerite
Erionite*	Melanophlogite	Tvedalite[a,b]
Faujasite*	Merlinoite	Wairakite*
Ferrierite*	Mesolite*	Weinebeneite[f]
Flörkeite	Montesommaite	Wenkite
Garronite*	Mordenite*	Wilhendersonite
Gaultite[c]	Mutinaite	Yugawaralite

notes:
* = zeolites recorded in the British Isles
a = beryllosilicates
b = structure not yet known
c = zincosilicate
d = beryllophosphate
e = berylloaluminosilicate
f = beryllophosphate

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Figure 2 Thomsonite, Monte Somma, Vesuvius, Naples, Italy (photo: Volker Betz http//www.iza-online.org/natural/).

Figure 3 Paulingite, Vinarickó, Hore, Nr. Kladno, Czech Republic (photo: Volker Betz (http//www.iza-online.org/natural/).

Figure 4 Levyne, Oderwiddeerrsheim, Vogelsberg, Hessen, Germany (photo: Volker Betz (www.iza-online.org/natural/).

Figure 5 Gmelinite, Avdello, Larnaka, Cyprus (photo: Volker Betz (http//www.iza-online.org/natural/).

Figure 6 Erionite, Agate Beach, Lincoln Co., Oregon, USA (http//www.iza-online.org/natural/).

Figure 7 Scolecite, Ósfjall, Breiödalsvik, Suóur-Múlasýsla, Iceland (photo: Volker Betz (www.iza-online.org/natural/).

species have yet to be found naturally. All recognised zeolite frameworks have been allocated a three-letter code by the IMA Structure Commission on Zeolites (Coombs *et al.* 1998)

Updated listing of all recognised frameworks can be found on the website of the International Zeolite Association (Ch. Baerlocher and L .B. McCusker, Database of Zeolite Structures: http://www.iza-structure.org/databases/). Table 1, based on this database, is confined to naturally occurring species. Note that this website also includes felspathoid minerals, because their frameworks are closely related to those of zeolites.

Coombs and his co-authors (1998) advise that some zeolites can be redefined as members of families that have the same framework, but differ in cation and/or in Si/Al content. For example, the gmelinite family has three members: gmelinite-Na, gmelinite-Ca and gmelinite-K. Other families are recognised based on the parent minerals: brewsterite, chabazite, clinoptilolite, dachiardite, erionite, faujasite,ferrierite, heulandite, levyne, paulingite, phillipsite and stilbite.

Industrial importance of zeolites

Natural species

Clinoptilolite is the most abundant natural zeolite and is mined in at least 16 countries from deposits of 'sedimentary' genesis (*see below*) (Fig. 8). In the UK the major use of this mineral makes use of its ability to scavenge 'ammonia' from wastewater, principally as a substitute for sand in swimming pool filters, but more recently for treatment of leachates from landfill sites. It is also available in garden centres as a soil amendment, and it is used to improve golf greens. Similar uses are common in other countries and clinoptilolite is widely used, for example, as food supplements for pigs, horses, poultry and even racing camels!

Figure 8 Zeolite mining in Zululand, RSA (Mineral Supplies International, with permission).

Another valuable cation exchange property of clinoptilolite is that it has the ability to selectively remove caesium and strontium radioisotopes from nuclear wastes. Material from the Death Valley Junction deposit near the Grand Canyon has been used highly successful in reducing isotope releases into the Irish Sea arising from the SIXEP process at Sellafield Ltd. This zeolite also found wide use in the clean-up following the Chernobyl disaster, including some 0.5 million tones imported from the Bulgarian Beli Plast deposit; these were dropped from the air in the initial entombment of the failed reactor. More recently clinoptilolite has been employed at Fukushima.

Chabazite is mined commercially in Bowie, Arizona, and used as a cheap gas desiccant/deodoriser as well as in the Three Mile Island clean-up and at the Oak Ridge Nuclear Labs in the USA. Ferrierite and mordenite have also been mined commercially from sedimentary deposits.

Zeolitic tuffs (chabazite/phillipsite) have been used in Italy as lightweight building materials since Roman times, and their worldwide use in building stone accounts for an estimated 100,000 tons per annum (tpa).

More details of these, and other uses, can be found in a publication from the Mineralogical Society of America (Bish and Ming 2001).

Synthetic species

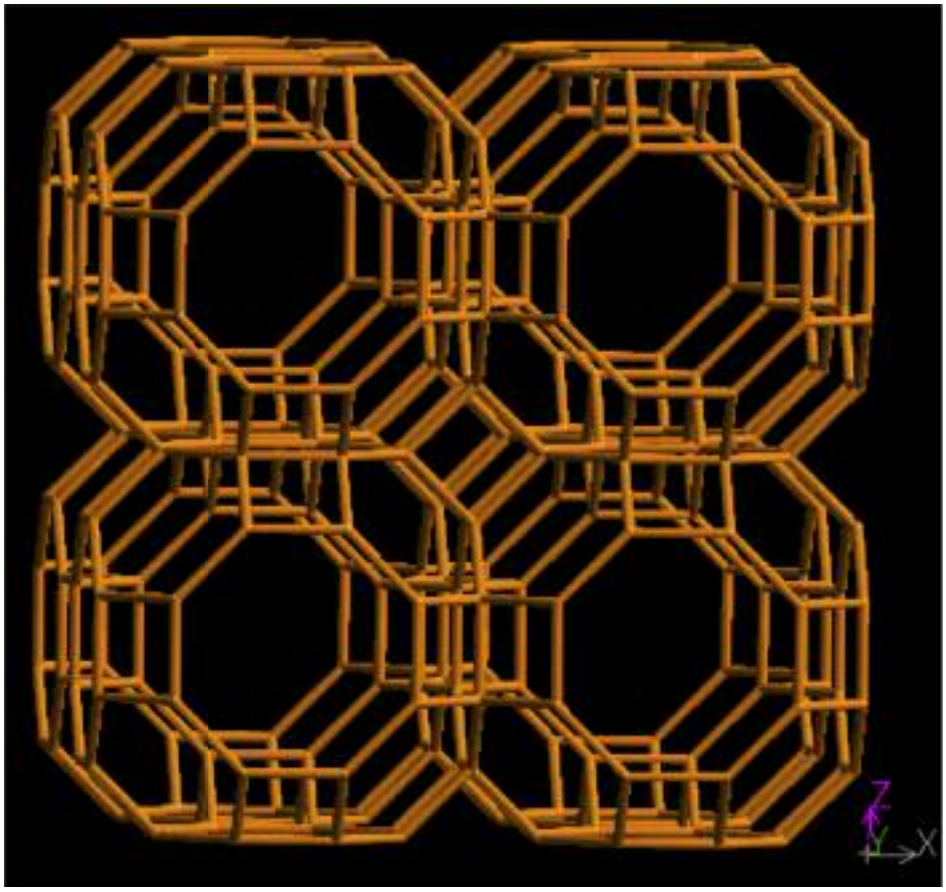
The first commercially successful synthetic zeolite was zeolite A, manufactured by the Linde Division of Union Carbide in the early 1950s. It has no known natural analogue.

It is present in most chemical laboratories in the world as 'Molecular Sieve 4A', for use to dry gases and liquids. Its desiccant properties are used in all refrigerator drying cartridges and in train air brakes. On a larger scale Molecular Sieve 4A is used to dry industrial gases, air lines and natural gas. The name 'molecular sieve' comes from the ability of this (and other synthetic species) to separate molecules, in both liquid and gaseous form, on the basis of their molecular size. The '4A' represents the 4 Ångstroms restrictive dimension of the access port, a ring of eight oxygen atoms to the A structure (Fig. 9).

This property facilitates the separation of straight-chain from branched-chain hydrocarbons, a property well used in the petrochemical and fuels industry. It also accounts for the use of Molecular Sieve 4A to remove unpleasant nitrogen- and sulphur-containing molecules from natural gas (a process known as 'sweetening').

The most valuable industrial use of zeolites came with the discovery that zeolite Y, a synthetic analogue of faujasite, is a powerful catalyst able to 'crack' the long-chain hydrocarbons in crude oil to the shorter lengths suited to useful fuels. Virtually all fuels used today at sometime come into contact with a zeolite catalyst. Prominent among them are those with a high per

Figure 9 The framework of synthetic zeolite A, each line represents an oxygen and each junction is the site of a silicon or aluminium. The diameter of the rings of 8 oxygens is 4Å (Ch. Baerlocher and L. B. McCusker, 'Database of Zeolite Structures: http://www.iza-structure.org/databases/).



cent of silica in their frameworks ('high silica zeolites') such as the zeolite ZSM-5, later discovered as a natural species in Antarctica — mutinaite — long after its first laboratory synthesis. Synthetic zeolites find numerous other uses as catalysts in the petrochemical, and other, industries.

Finally, the largest commercial bulk usage of synthetic zeolite species is their addition to domestic and commercial detergents, which accounts for more than one million tpa. A typical supermarket pack will contain *c.* 17% of either zeolite A or the synthetic gismondine zeolite MAP. Their purpose is to act as water softeners ('builders') as replacements for the environmentally damaging phosphates.

There are many other applications of zeolites and those readers who wish to find out more about these unique materials are referred to recent reviews (e.g. Čjeka *et al.* 2007).

The genesis of zeolites in nature

Zeolites can be formed in a variety of water-rich environments, usually at low pressures and at temperatures below 300°C (Fig. 10). They can have hydrothermal, igneous, metamorphic or sedimentary origins. The species, and amount of, a zeolite formed varies primarily as a function of texture and composition of the host rocks; the composition of interstitial water; and age and temperature.

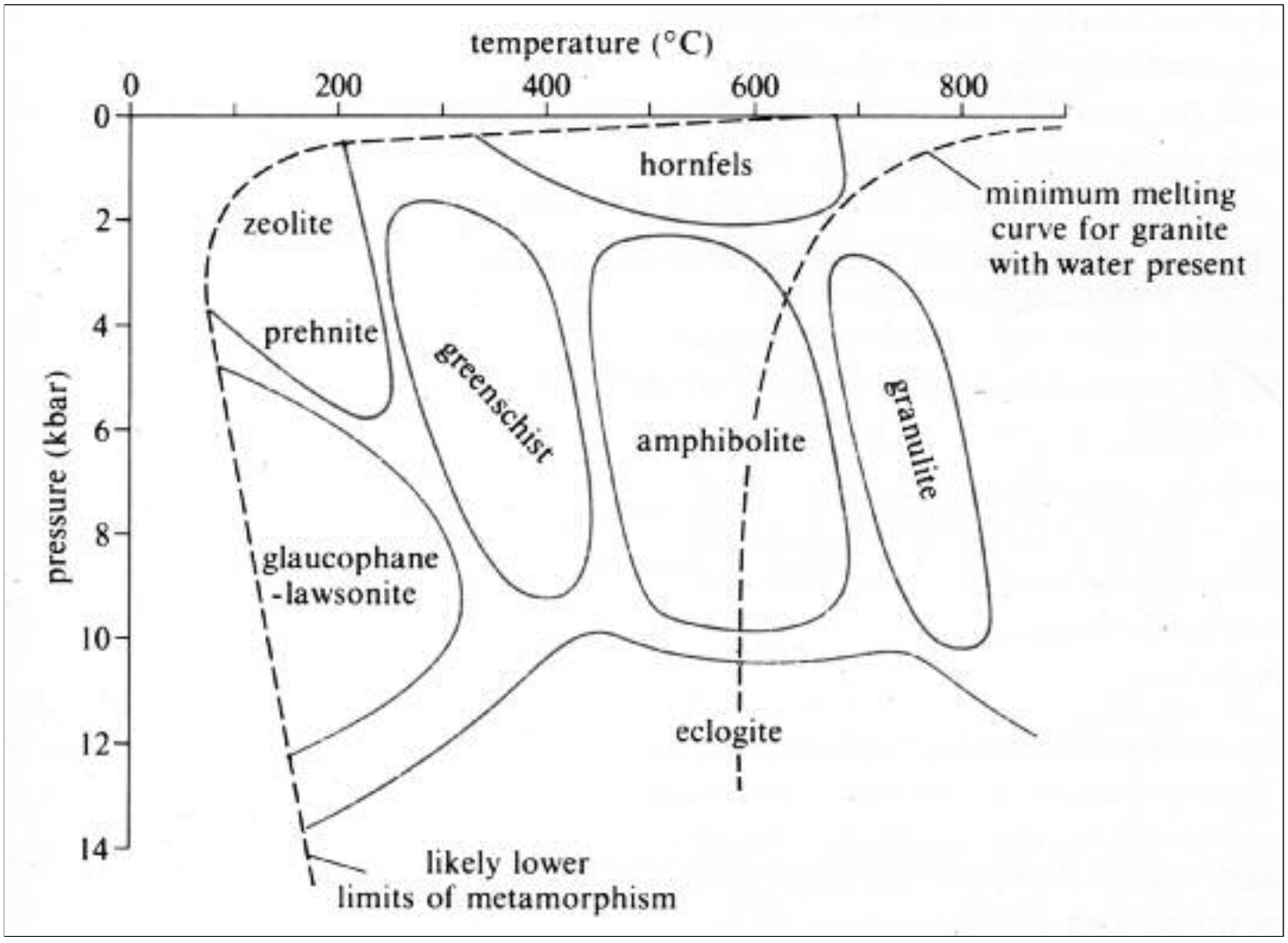
The minute zeolite crystals that form at lower temperatures (4°C–40°C) in sedimentary deposits have required long crystallisation periods, sometimes millions of years. Often these deposits consist of volcanic debris as a common precursor material for zeolite formation. Large zeolite crystals form in igneous or hydrothermal environments, as these provide the higher temperatures needed to produce the larger crystals (40°C–300°C, rarely higher). Elevated temperatures speed up chemical reactions, so large crystals of the same minerals can crystallise in a few months or years. Any part of the world that has experienced the necessary conditions for zeolite genesis is a likely source of zeolites or zeolite-bearing rocks, subject to age and subsequent geological events.

Types of zeolite genesis

Lithogenesis (in rocks)

Lithogenic zeolites can form *in situ* from both volcanic materials, e.g. clinoptilolite in the Catahoula Formation, South Texas, USA, and non-volcanic materials, e.g. laumontite in the San Gabriel Mountains, South California, USA. They also form in materials of Aeolian or alluvial deposition, e.g. clinoptilolite in the Aeolian Norwood Tuff Formation, Northern Utah, USA and alluvial clinoptilolite in the Rio Grande River, USA/Mexico.

Figure 10 Zeolite formation in relation to other silicate minerals (from Open University Course S236 Geology, Block 3 Internal Processes*, 109, [with permission]).*



Pedogenesis (in soils)
Pedogenic zeolites are found in saline, alkaline, soils of volcanic parent materials. Examples include the widespread zeolites in the Olduvai Gorge, Tanzania and in the Wright Valley, Antarctica. They also form in saline, alkaline soils of non-volcanic parent materials. An example of this is the analcime in the San Joaquim Valley, California, USA.

Genesis by chemical weathering after evapotranspiration
Zeolites can also be formed by chemical weathering in arid or semi-arid regions, in soils or in land surfaces rich in sodium carbonate or bicarbonate. If suitable precursor materials are present, such as volcanic tuffs or clays, and if the pH is high enough due to the concentration of sodium bicarbonate by evapotranspiration, then zeolites will readily form.

Diagenesis in hydrologically closed systems
Saline and alkaline lakes are hydrologically closed systems. Such lakes are found in arid or semi-arid regions and are formed in closed basins. Water leaves the system only through evaporation, producing lake waters of high pH (*c*. 9.5) due to high concentrations of dissolved carbonates and bicarbonates. Zeolites form by diagenesis from reactive materials deposited in these lakes. The precursor materials are often wind blown volcanic material, such as vitric tuffs. Other suitable materials included biogenic silica from skeletons of radiolarians and forams, clays, plagioclase feldspar and forms of quartz.

Worldwide there are abundant sedimentary deposits of this origin and they form the core of the current interest in mining natural zeolites for commercial use. Clinoptilolite is the main zeolite to be mined.

Genesis in hydrologically open systems
In hydrologically open systems zeolites are precipitated when meteoric water flows through porous beds of lava, ash or tuff. This water percolates through thick layers of material such as palagonite tuff, a pyroclastic rock rich in reactive glass. As the water penetrates to greater depth it becomes more alkaline and saline resulting in a vertical zonation characteristic of the genesis. Near the surface, glass is altered to clay minerals; a little deeper, to zeolites such as clinoptilolite, chabazite and phillipsite; and deeper still, to analcime. At the lowest level, alteration is to alkali-feldspars. In open systems usually only one or two species of zeolites are formed.

Genesis in geoautoclaves
This type of genesis has been proposed to explain the occurrence of zeolites in rocks such as ignimbrites (ash flow deposits), where the ignimbrite has been homogeneously zeolitised while associated tuffs above and below it totally lack zeolites, despite having the same chemical composition. Open-system diagenesis cannot account for this and it has been proposed that a geoautoclave system exists in which zeolitisation is related to the high temperatures (possibly up to 200°C) and high vapour pressures existing in the rock unit immediately after its deposition.

Another hypothesis is that some difference in the nature of the glass, such as the grain size of the glass fragments and consequently their different surface/volume ratios, could be the key to explain why one tuff is zeolitised and another is not. This type of genesis needs further study but possible examples include the Roque Nublo ignimbrites of Gran Canary Island, Yucca Mountain in Nevada, and the Rhodope Mountains in Bulgaria and Greece.

Burial diagenesis and very low-grade metamorphism
With this type of genesis, zones differing with increasing depth are recognised and are similar to those for the hydrologically open system of diagenesis, except that in burial diagenesis the zones are thicker (100m to 1,000m instead of only *c.* 10m). If any hydrothermal influence is present the zones are thinner and do not have horizontal boundaries. A characteristic of burial diagenesis is that only a limited number of zeolite species crystallise. These include clinoptilolite, mordenite and laumontite.

It is likely that some of the minute-sized zeolite crystals of the clinoptilolite-heulandite minerals found in south-east England are examples of burial diagenesis and low grade metamorphism. They have been recorded from the Corallian Beds of Jurassic age and were found in sediments taken from borehole cores from beneath the Harwell Research site in South Oxfordshire. They have also been found in Fuller's Earth deposits of Cretaceous age found in Surrey and Bedfordshire, as well as in the Thanet Sands of Palaeocene age in Kent.

Diagenesis in shallow marine sediments
There is a broad variety of these sediments. In general they are volcaniclastic and could have been zeolitised above or below sea level; and may, or may not, now crop out on land. The formation of zeolites in hyaloclastics is currently thought to occur below sea level when seawater reacts with fine-grained basaltic, glassy fragments. Small glass fragments are usually produced by explosive volcanic activity in a shallow sea (hyaloclastites), but they can also result from the fragmentation of a small lava flow as it enters the sea; a large lava flow would not be chilled rapidly enough. The first product of the glass and seawater reaction is that of palagonitisation, i.e. the hydration of glass. The final products are clay minerals such as smectites, zeolites (mainly phillipsite, but also chabazite and gonnardite), calcite and iron hydroxides. An example of this is in the pillow lavas of the Pleistocene Honolulu Group, Ohaha, Hawaii.

Diagenesis in deep-sea sediments
The largest volumes of zeolites found are contained in sea-flow sediments. Of these sediments it has been estimated that, in the Pacific, clinoptilolite averages 2% and phillipsite 1.5%; analcime is next in abundance.

Clinoptilolite occurs widely in both tuffaceous and non-tuffaceous sediments, and particularly in carbonates. It is most common at depths greater than 100m and increases towards a maximum in Upper Cretaceous sediments. Phillipsite is generally associated with volcanic materials, mainly basaltic. It is found principally in Miocene and younger sediments at burial depths of less than 100m, but it can occur in sediments as old as Cretaceous. Analcime content increases with age and is most abundant in the oldest sediments sampled, which are of late Jurassic age. The analcime can be directly associated with glass. Glass alters only slowly in deep-sea sediments and unaltered glass as old as the Cretaceous has been found.

Other zeolites found in deep-sea sediments include natrolite, heulandite, harmotome, erionite and mordenite. The last two types occur rarely. The world's deep-sea area is divided into

different ocean regions and their marginal seas. The distribution of deep-sea zeolites is distinct in each region. For example, clinoptilolite is more common than phillipsite in the Atlantic. Parts of the British Isles are bordered by the Atlantic and the seas around the coastline form extensions of this ocean. Clinoptilolite comprises 88% of the zeolites present in the Atlantic sediments.

Hydrothermal genesis in geothermal fields
Zeolites are often present in sediments found in areas of geothermal activity, such as Yellowstone National Park, Wyoming. This is especially so if some tuffs, with glassy components, are included among the sediments. The zeolite mineralisation occurs when hot water, 50°C–150°C, dissolves silicates and glass in the deepest and hottest parts of the sediments, and then deposits zeolites throughout the whole of the sequence. This kind of genesis characteristically produces a large variety of zeolite species and can also feature a continuous transition to burial diagenesis.

Hydrothermal genesis in ore deposits
Ore deposits that have been formed by crystallisation from hot solutions (150°C–400°C) under pressure in veins or in porous rocks sometimes contain zeolites. In the British Isles there are a number of examples of zeolites formed in these conditions, such as in the North Pennines ore field.

Late hydrothermal deposition of pegmatites
The temperature of crystallisation of pegmatites is too high for the formation of zeolites. However, some pegmatites do contain zeolites. These were probably formed at lower temperatures as a result of late hydrothermal deposition on previously crystallised pegmatite minerals. Examples can be seen in de-silicated pegmatites at Hrubsice, Western Moravia and in the Tonco pegmatite, Bernic Lake, south-east Manitoba, Canada.

Hydrothermal veins and geodes in feldspathic rocks
Zeolites can also crystallise from the reaction of water with the minerals in feldspar-bearing rock such as granite, granodiorite, gneiss, amphibolite and similar coarse-grained non-volcanic, feldspar-bearing rocks. Upcoming hydrothermal fluids can cause a dissolution-deposition process to occur, and this may result in the crystallisation of zeolites in any veins and geodes present within these rocks. An example of this is the Takinoue geothermal area of northern Honshu, Japan.

The genesis of zeolites in geodes and vugs of basalts
Many of the finest zeolite specimens are found on the walls of geodes and vugs formed in basaltic lavas. One hypothesis is that upcoming thermal water, possibly of magmatic origin, dissolves the glassy components in the hottest part of a basalt flow. Zeolites then crystallise out of the thermal water in regions at lower temperatures.

Another hypothesis is similar to that for open-system diagenesis. Here meteoric water percolates through fractures in the basalt flow, the basalt glass is dissolved and zeolites are deposited in regions close to the dissolution area. In both processes the temperatures involved could be similar, as water can be considered hydrothermal even below 100°C; and diagenesis can also occur above 100°C if it is at the appropriate depth. In addition the distance between the sites of dissolution and deposition is not necessarily zero in an open-system diagenesis, or very great in hydrothermal genesis. There can therefore be a continuum between the two models, with no clear-cut distinctions between them.

The hydrothermal hypothesis could be more likely for many of the occurrences because large zeolite crystals, apart from basalt geodes, always have a hydrothermal genesis ,while diagenetic zeolites are usually microcrystalline (*c*. 15μm).

Microcrystalline zeolites from sediments have a different average chemical composition from those found in vugs. In addition, the number of zeolite species formed by diagenesis is limited, whereas hydrothermal genesis is a possible source for all zeolite species, and in the vugs of basalts a variety of zeolites do occur. The British Isles have fine examples of zeolites formed by this type of genesis. In fact it gives rise to a very large percentage of the zeolites found in Scotland and Northern Ireland. A variety of zeolites such as analcime, chabazite, stilbite and thomsonite can commonly be found in the Tertiary lavas of Mull, northern Skye and Antrim, and also in the Carboniferous lavas of the Midland Valley of Scotland.

Magmatic genesis
A magmatic genesis for zeolites is rare and analcime has been cited as the only zeolite that has been proved to have originated in this way. The only known example of a truly magmatic analcime is in lavas of minette composition in the Colima volcanic complex in Mexico, an occurrence proved by oxygen isotope analysis.

In its formation in magma, analcime can only crystallise from a silicate melt in a narrow temperature and pressure range. Basically the analcime of lavas is sodium exchanged leucite. However, note that in the IMM 1998 reclassification of zeolite species, leucite is now included as a zeolite (Coombs *et al.* 1998), as listed in Table 1.

Zeolites in skarn deposits
There have been several records of skarn zeolites. Examples are those of thomsonite, natrolite, mesolite, scolecite, chabazite, heulandite, stilbite, stellerite and analcime found in skarn iron ore deposits at Arendal, Norway. They are thought to be derived from the older skarn minerals such as scapolite.

Impact crater occurrences
Zeolites have been discovered in suevite and lake sediments of the Noordlinger Ries impact crater in Bavaria, West Germany. Microcavities of suevite are filled by analcime, clinoptilolite, phillipsite, erionite and harmotome. Recent research at The Open

Figure 11 Lava beds at Gjogl, Eysturoy, Faroes (photo: A. Dyer).

University confirms the presence of clinoptilolite in the Boltysh impact crater in the Ukraine (personal communication: Felicity Williams, The Open University).

Zeolite zones

In some types of zeolite genesis, different species will form in specific zones. The order of species formed is dependent on temperature. The zones can occur across a region through contact metamorphism, or vertically through layers of lava flows, such as in the Faroes Islands lava flows shown in Figure 11 *(opposite)*.

In the British Isles the formation of regional zones of preferred zeolites has been recorded in the Tertiary basalt lavas of northern Ireland (Walker 1970) and also those of northern and central Skye in Scotland (King 1976).

Age of zeolites

In nature zeolites have a finite lifetime because they eventually alter to feldspars. Consequently they are not likely to be found in rocks older than the Carboniferous Period (359–299Ma) unless there has been some subsequent geological event that provided the conditions for zeolite formation within older host rocks. Most of the world's zeolites are found in rocks of Jurassic age (199–145Ma) or younger.

What Next?

In a second article we will discuss zeolite species found in Great Britain. We will include descriptions of the more common zeolites found here, how to recognise them and where to find them.

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